

High-resolution geophysical observations and modeling of the cryosphere using distributed optical fiber sensing

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Abstract

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Earth and Space Sciences

Glaciers and ice sheets are projected to rapidly decline in volume and contribute significantly to sea level rise in the coming centuries due to anthropogenic climate warming. The risk to fresh water availability in coastal infrastructure are great; however, our understanding of how glacier melt and ice discharge vary over space and time is limited. Current conventional instrumentation for sensing glacier and ice sheet change due to melt or ice discharge can be classified into two categories: remote sensing and in situ instrumentation. Remote sensing platforms have the benefit of a large aperture, but often are limited in revisit times and resolution. In situ instrumentation, however, can make continuous measurements of glacier change limited only by the labor to deploy a network of sensors and the sensor robustness. This motivates the search for novel methods for sensing of glacier melt and flow that bridges the gap between aperture and spatiotemporal resolution. Here, I leverage distributed optical fiber sensing, namely distributed acoustic sensing and distributed temperature sensing. Both systems allow for meter scale sampling of the acoustic or temperature fields continuously along a fiber optic cable with collection duration times of months if not longer. In the following chapters, I utilize distributed optical fiber sensing techniques to make novel observations of glacier melt and ice sheet flow.

Using a fiber cable deployed along the entire flow line of Rhonegletscher in Switzerland, I calculate the spatiotemporal variability of the supraglacial seismo-acoustic wave field amplitude. I first show that

the supraglacial seismo-acoustic wave field is a function of the amount of meltwater runoff and use this relationship to infer proglacial discharge. Following from this work, I deployed fiber optic cables in boreholes drilled in the Allan Hills Blue Ice Area, Antarctica. Here, I show how borehole temperatures record climate in the blue ice area and model ice flow that reveals mechanisms for preserving ancient ice. Finally, I perform a passive seismic survey at the Eastwind Glacier grounding line, Antarctica and reveal passive seismic reflections from the grounding line. I use the seismic reflections to recover a reflection coefficient and infer grounding line material properties. Through this work, I demonstrate the power of distributed optical fiber sensing systems for seismic and temperature sensing over large spatial and temporal baselines on glaciers. Furthermore, I add new insights into how the surface of glaciers change throughout a melt season, different mechanisms for preserving old ice that operate in a blue ice area, and Antarctic grounding line properties using observations of never before seen passive seismic reflections. This work opens the door to the possibility of monitoring highly dynamic and sensitive glacier environments as glaciers begin to respond to climate change.

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Dedication

to Mom, Dad, Brittny, and Christian

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Chapter 2

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Chapter 3

The GNSS data is available at <https://doi.org/10.15784/602026>. Borehole temperature data can be found at <https://doi.org/10.15784/602025>. The analytical solution is plotted using code available at <https://doi.org/10.5281/zenodo.18526907>. Previously unpublished $\delta^{38}\text{Ar}/^{36}\text{Ar}$, $\delta^{18}\text{O}$, and ^{40}Ar ice core ages are available at <https://doi.org/10.15784/602044>.

Chapter 4

Distributed acoustic sensing data can be provided at reasonable request to the author. Codes used to produce chapter figures can be found at https://github.com/johnmorganmanos/eastwind_v2

Chapter 1

Introduction

1.1 Background and Motivation

Glaciers and ice sheets have been features on Earth's surface through many parts of Earth's geologic history and the extent of glaciation has varied widely. At one end of the extreme, Earth experienced extensive glaciation during the Neoproterozoic ($\sim 750 - 550$ Ma) with ice cover likely reaching as far as the tropical latitudes [Hoffman et al., 1998]. Conversely, Earth was largely ice free during the early Eocene (52 Ma) when sea surface temperatures were as much as 8°C warmer on average [Zachos et al., 2001]. The present-day Antarctic Ice Sheet formed ~ 34 Ma [DeConto & Pollard, 2003] with possible glacial inception occurring in the high-elevation Transantarctic Mountain Range as early as ~ 60 Ma [Barr et al., 2022]. The Greenland Ice Sheet, on the other hand, formed at least 18 Ma [Thiede et al., 2011]. More recently, the onset of Northern Hemisphere glaciation (excluding Greenland) occurred around 3.6 Ma with a marked intensification occurring ~ 2.7 Ma [McClymont et al., 2023]. Since the most recent inception of glaciation, the glaciers and ice sheets have cyclically advanced in their area and retreated, forced by changes in the Earth's insolation due to variations in Earth's orbital cycles around the Sun [Jouzel et al., 2007; Imbrie et al., 1992]. In addition to this periodicity, there are rapid changes to climate and Earth ice cover that take place over less than a few hundred years [Dansgaard et al., 1989] as a result of the coupled ice-ocean-atmosphere interactions. Since the start of the Industrial Era (1850 AD), however, human activity has significantly influenced Earth's climate system beyond the expected natural variability [Calvin et al., 2023; Roe et al., 2021]. As a

consequence, glaciers and ice sheets worldwide have lost more than 267 Gt yr^{-1} since 2000 and this trend is accelerating [Hugonnet et al., 2021].

Under high emission scenarios, glaciers worldwide are expected to lose about 64 percent of their global total volume by the end of the century [Shannon et al., 2019]. The glacier volume loss is not uniform for all glaciated regions. The European Alps, for example, are expected to fair much worse and be nearly ice free by 2100 [Zekollari et al., 2019] whereas net mass gain over the East Antarctic Ice Sheet has roughly balanced mass loss across the Antarctic continent [Wang et al., 2021]. Nevertheless, global net ice mass loss will have dire consequences for human populations that depend on glacier meltwater for drinking, power, and irrigation in addition to threatening coastal populations as a result of subsequent sea level rise. While short term availability of fresh water resources will increase with increased glacier runoff, the demand for primary production that need large amounts of fresh water is increasing and will strain a diminishing water supply in the future [Mark et al., 2015].

Determining future glacier melt and dynamical response, however, is nontrivial. To predict future glacier melt and ice sheet discharge, numerical models are employed to solve equations describing glacier melt and flow (e.g. [Schoof, 2007]). Projections of glacier melt are highly sensitive to the type of melt model used [Gabbi et al., 2014]. Simple melt models that only depend on temperature (termed positive degree day models) are generally accurate at predicting melt, but the scaling factor can vary spatially and temporally leading to significant melt uncertainties [Braithwaite, 1995]. In these instances, it may be appropriate to use more complex models that consider the full energy and turbulent flux balance. Though full energy balance models more accurately predict glacier melt particularly over shorter time (hourly) and spatial (meters) scales, they still require assumptions regarding ice or snow surface characteristics and high resolution meteorological data [Hock, 2005]. In locations, such as Antarctica, where mass discharge is dominated by ice dynamical processes [Rignot et al., 2019], numerical glacier flow models are employed to quantify mass discharge.

Glacier flow models apply a range of approximations to the mass conservation and momentum equations (or Navier-Stokes equations) written as

$$\nabla \cdot \mathbf{u} = 0 \tag{1.1}$$

$$\frac{\partial}{\partial x}(2\mu \frac{\partial u}{\partial x}) + \frac{\partial}{\partial y}(\mu \frac{\partial u}{\partial y} + \mu \frac{\partial v}{\partial x}) + \frac{\partial}{\partial z}(\mu \frac{\partial u}{\partial z} + \mu \frac{\partial w}{\partial x}) - \frac{\partial p}{\partial x} = 0 \quad (1.2)$$

$$\frac{\partial}{\partial x}(\mu \frac{\partial v}{\partial x} + \mu \frac{\partial u}{\partial y}) + \frac{\partial}{\partial y}(2\mu \frac{\partial v}{\partial y}) + \frac{\partial}{\partial z}(\mu \frac{\partial v}{\partial z} + \mu \frac{\partial w}{\partial y}) - \frac{\partial p}{\partial y} = 0 \quad (1.3)$$

$$\frac{\partial}{\partial x}(\mu \frac{\partial w}{\partial x} + \mu \frac{\partial u}{\partial z}) + \frac{\partial}{\partial y}(\mu \frac{\partial w}{\partial y} + \mu \frac{\partial v}{\partial z}) + \frac{\partial}{\partial z}(2\mu \frac{\partial w}{\partial z}) - \frac{\partial p}{\partial z} = 0 \quad (1.4)$$

where $\mathbf{u}$ is the velocity vector (u, v, w) in the (x, y, z) directions, μ is the viscosity, and p is pressure. Simple ice flow models neglect the horizontal gradient components of the momentum equations as they are orders of magnitude smaller than the vertical gradient component [Hutter, 2017]. These models are good at approximating ice sheet flow, but quickly break down where there is steep bedrock topography [Le Meur et al., 2004]. Here, all components of the momentum equation must be considered (called a full Stokes model). The viscosity μ is often parameterized using a constitutive relationship that describes the deformation of ice to an applied stress—i.e. the power law relationship called Glen’s Flow Law:

$$\dot{\epsilon} = A(T)\sigma^n \quad (1.5)$$

where $\dot{\epsilon}$ is strain rate, A is the prefactor which depends on temperature, σ is stress, and $n \approx 3$ is the flow law exponent [Glen, 1955]. Nye [1957] generalized this relationship to all deviatoric stress components such that

$$\tau_{ij} = 2\mu\dot{\epsilon}_{ij} \quad (1.6)$$

where τ_{ij} and $\dot{\epsilon}_{ij}$ are the shear stress and strain rate tensors, respectively.

To solve these equations, we can use the *finite element* method whereby we discretize the domain into *elements* that comprise a mesh of the domain and solve the weak form of the equations above (eqs. (1.1) to (1.4)) using linear algebraic solvers implemented in Python software (e.g. *Firedrake* [Ham et al., 2023]). All models regardless of simplicity, however, rely on appropriate definition of the boundary conditions and observations to either constrain the model or to validate its accuracy. By improving the quality of our

observations, better models can be produced.

Advancements in laboratory and field measurements demonstrated the need to improve glacier flow laws and models to account for different glacier bed conditions such as deforming till [Zoet & Iverson, 2020] and basal hydrology [Bartholomew et al., 2010]. The spatial and temporal scales that are important to consider for characterizing glacier melt and flow are wide and varied—from the grain scale (e.g., ice rheological properties) to the continent scale (e.g. ice sheet surface mass balance). For example, the frictional properties at the glacier bed, the force resisting glacier ice flow, can vary over meters [Gräff et al., 2021] to tens of kilometers [Barcheck et al., 2019]. Remote sensing is capable of illuminating glacier processes over the ice stream scale for example, but it cannot capture all dynamical processes important for driving glacier flow necessitating *in situ* instrumentation [Olinger et al., 2024]. Detection of englacial or subglacial processes are fundamentally obscured from remote sensing platforms. Using *in situ* seismological instrumentation to record ambient noise and icequakes (the brittle fracture of ice and coseismic glacial slip at the bed) [Ekström et al., 2003] can help illuminate glacial melt, slip, and flow processes [Podolskiy & Walter, 2016; Fichtner et al., 2025]. In this dissertation, I apply novel methods in seismological sensing of the cryosphere using distributed optical fiber sensing (DOFS) to explore challenging questions related to glacier motion and melt where traditional instrumentation has historically failed. The culmination of this work amounts to a significant step forward in both the techniques of DOFS deployments in glacier environments and our understanding of the spatial and temporal variability of glacier response to its environment.

1.2 Methods: Distributed Optical Fiber Sensing (DOFS)

Distributed optical fiber sensing (DOFS) has gained significant popularity in industrial infrastructure monitoring and more recently for applications in geophysical research (see Fig. 1.1) [Zhan, 2019]. The sensing element, fiber optic cables, are ubiquitous in today's environment: they span the ocean floor delivering data from one continent to another, cross urban centers, and up to high mountain passes meeting internet connectivity needs. Previously deployed telecommunication fiber cables not presently in use, called "dark" fiber, can be leveraged to create a well coupled seismic array where deploying individual seismic nodes is impractical or impossible [Fernández-Ruiz et al., 2020]. In geophysical research, using fiber optic cables is attractive as they are relatively cheap (<\$700 for 1 km of armored cable) and less susceptible to losses in

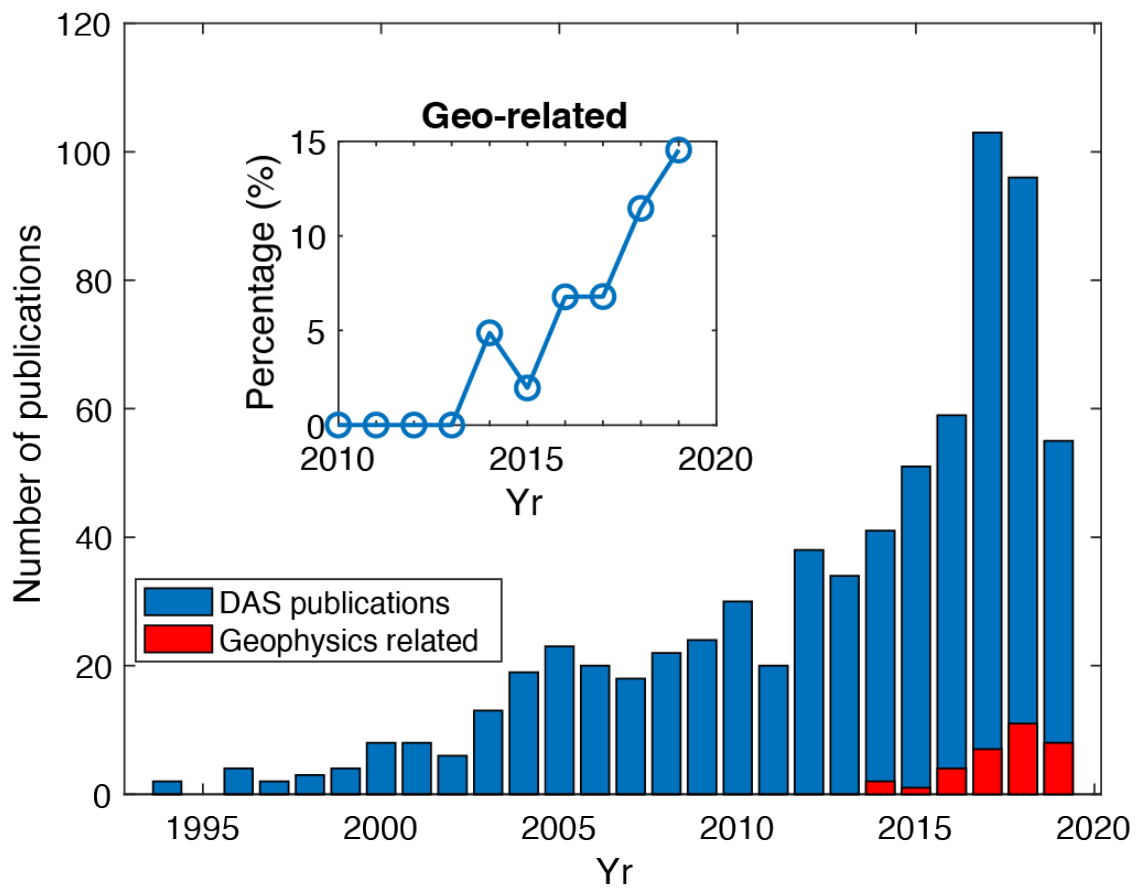


Figure 1.1: The number of research publications referencing distributed acoustic sensing (DAS). Figure from [Zhan, 2019]

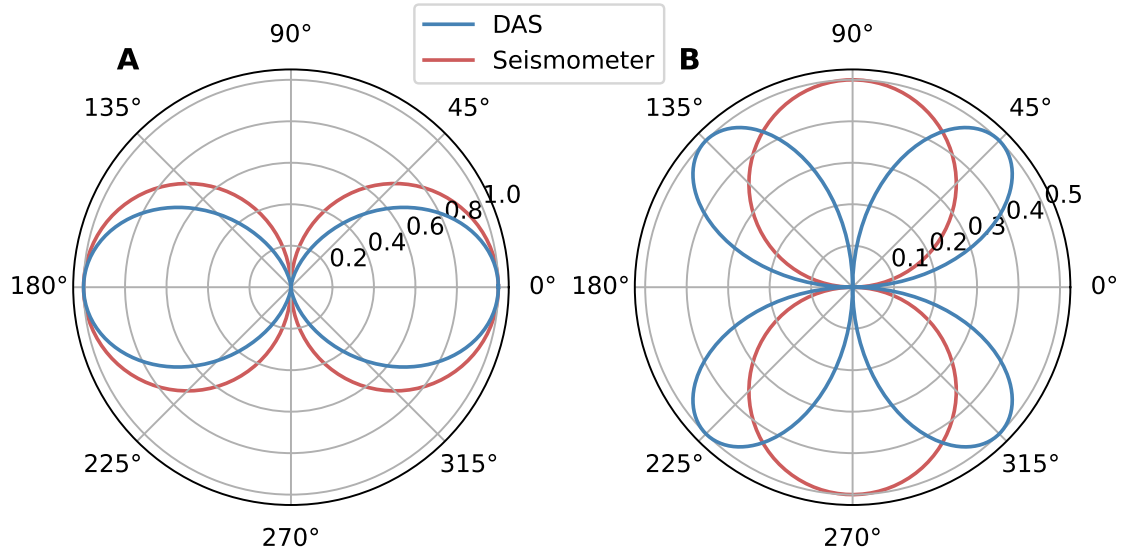


Figure 1.2: (a) Amplitude response of a DAS longitudinal strainmeter and seismometer to a P (acoustic) wave. (b) Similar to (a), but a S (shear) wave. Figure adapted from [Benioff, 1935] and Zhan [2019]

data quality due to surface melt out or other environmental disturbances. Yet, utilizing DOFS to investigate glacier melt and glacier dynamical processes is in its infancy and the full capabilities are still being explored. In this dissertation, I use two forms of distributed optical fiber sensing: distributed acoustic sensing (DAS) and distributed temperature sensing (DTS).

1.2.1 Distributed Acoustic Sensing (DAS)

Distributed acoustic sensing (DAS) interrogators pulse laser light at high rates into a single-mode fiber optic cable. The pulsed light scatters elastically off of imperfections in the crystal matrix of the silicate glass within the core and some of the light returns back to the photons detectors in the interrogator. The scattered light is in the same frequency as the incident light and is called Rayleigh scattering. The photon detector measures the phase of the returned Rayleigh backscatter, which is proportional to fiber strain, over

continuous sections of the fiber optic cable called the "gauge length" and assigns the value to a subset length called the "channel" with location determined by the two way travel time of the light. DAS does this repeatedly at high rates along the optical path length effectively turning a fiber optic cable into a dense array of one-dimensional strainmeters [Lindsey & Martin, 2021]. The gauge length may be and is often greater in size (distance) than the channel spacing as this generally improves the signal-to-noise ratio (SNR) [Dean et al., 2017]. DAS interrogators have a spatial sampling resolution of 1-10 m and have a broad bandwidth from 0.01 Hz to ~ 30 kHz. DAS has been used to observe a range of geophysical processes from ocean gravity waves [Williams et al., 2022], teleseismic earthquakes [Williams et al., 2019], calving [Gräff et al., 2025], and icequakes [Hudson et al., 2021; Fichtner et al., 2026]. Because DAS is only sensitive to longitudinal strain in the direction of the fiber optic cable, it has an amplitude response proportional to $\cos^2 \theta$ for acoustic (P) waves and $\sin \theta \cos \theta$ for shear (S) waves whereas conventional seismometers have responses proportional to $\cos \theta$ and $\sin \theta$ for P waves and S waves, respectively (Fig. 1.2). For this reason, more seismic events may be detected and source locations can be better constrained by using a two-dimensional deployment array configuration (i.e., including turns in the cable) over a purely linear array [Hudson et al., 2021].

1.2.2 Distributed Temperature Sensing (DTS)

Like DAS, distributed temperature sensing (DTS) interrogators utilize multi-mode fiber optic cables for the instrument sensing element; however, the Raman backscatter in the Stokes and anti-Stokes frequencies is used to sample the temperature field. Raman scatter in a fiber optic cable is a result of the interaction between the pulsed light by the DTS interrogator and molecular lattice oscillations in the fiber optic cable [Ukil et al., 2012]. If there is energy absorption into the lattice, the molecule emits light at a lower frequency (Stokes wavelength). If the molecule drops to a lower energy state due to the interaction with pulsed light, the molecule emits light in a higher frequency (anti-Stokes wavelength). The ratio between the anti-Stokes and Stokes emissions can be used to quantify the temperature of the fiber continuously along the optical path. The temperature measurement can be affected by the intrinsic differential attenuation of the light in the fiber. This can be corrected for using double-ended configuration (commonly in a "U-bend") where light pulses are alternated between one end of the cable (forward) and the other end (reverse) to retrieve two independent

attenuation measurements at the same point of the fiber optic cable. After attenuation correction, temperature precision of $\sim 0.01^{\circ}\text{C}$ are possible with spatial resolution of ~ 1 m [Selker et al., 2006]. DTS has been successfully deployed to study soil moisture [Steele-Dunne et al., 2010], ocean temperature oscillations [Sinnott et al., 2020], and warm water infiltration in a sub-ice-shelf cavity [Tyler et al., 2013].

1.3 Dissertation Organization

The following chapters of this dissertation are organized into three distinct sections that encompass a selection of the prepared and published work I have led during the course of my studies. Upon first glance, the chapters in this dissertation may appear disparate, as the study sites are each unique and each chapter focused on a different dynamical process; however, using advancements in computing and high-resolution sampling techniques, each chapter explores glacier dynamics in the extremes of the cryospheric domain. Figure 1.3 shows an overview map which highlights the study sites for each chapter.

In Chapter 2, I investigate the supraglacial soundscape of a rapidly melting mountain glacier in the Swiss Alps. We demonstrate the power of machine learning models to learn on DAS recorded ambient noise of turbulent water flow and predict proglacial discharge. In Chapter 3, I create a thermo-mechanically coupled full Stokes ice flow model of the Allan Hills, Antarctica Blue Ice Area (BIA) where steep bed topography requires the use of higher order (e.g. full Stokes) models to accurately describe the present day flow field. Here, I propose two mechanisms for preserving old ice in Antarctic BIAs: 1) convergence and subsequent stagnation flow upstream of exposed nunataks and 2) advection of old ice from elsewhere. The model is validated by Global Navigation Satellite System (GNSS) surface velocity measurements and DTS borehole temperature profile measurements. In Chapter 4, I record passive seismic data across the Eastwind Glacier, Antarctica grounding line using DAS. All detected events are located using a Markov chain Monte Carlo sampling of the parameter space and correlations between the grounding line geometry, tidal cycles, and icequake frequency are revealed. I finish the dissertation with a short summary of the work presented here and a perspective on possible future directions in modeling and observing glacier dynamics at high resolution. I finish by discussing some perspectives on possible future avenues of research.

Each chapter in this dissertation, of which I am lead author, has either been published in an academic journal and includes an accompanying citation, has been submitted for peer review prior to publication, or

is in preparation for submission as indicated in each chapter's preamble. However, each chapter includes significant contributions from co-authors for which I am grateful.



Figure 1.3: Overview map highlighting the locations of the studies presented in each chapter of this dissertation.

Chapter 2

DAS to Discharge: Using Distributed Acoustic Sensing (DAS) to infer glacier runoff

Originally published in the Journal of Glaciology [Manos et al., 2024]

2.1 Chapter Abstract

Observations of glacier melt and runoff are of fundamental interest in the study of glaciers and their interactions with their environment. Considerable recent interest has developed around distributed acoustic sensing (DAS), a sensing technique which utilizes Rayleigh backscatter in fiber optic cables to measure the seismo-acoustic wavefield in high spatial and temporal resolution. Here, we present data from a month-long, 9 km DAS deployment extending through the ablation and accumulation zones on Rhonegletscher, Switzerland, during the 2020 melt season. While testing several types of machine learning (ML) models, we establish a regression problem, using the DAS data as the dependent variable, to infer the glacier discharge observed at a proglacial stream gauge. We also compare two predictive models that only depend on meteorological station data. We find that the seismo-acoustic wavefield recorded by DAS can be utilized to infer proglacial discharge. Models using DAS data outperform the two models trained on meteorological data with mean

absolute errors (MAE) of 0.64 m³/s, 2.25 m³/s, and 2.72 m³/s, respectively. This study demonstrates the ability of *in situ* glacier DAS to be used for quantifying proglacial discharge and points the way to a new approach to measuring glacier runoff.

2.2 Introduction

Glaciers are an important yet diminishing reservoir of freshwater for communities and ecosystems [Casassa et al., 2009]. In the European Alps, for example, modeled future trends indicate a large reduction or disappearance of glaciers on decadal timescales due to climate change [Haeberli et al., 2007; Linsbauer et al., 2013; Zekollari et al., 2019]. Glaciated catchments provide a river discharge buffering mechanism, particularly important during the dry season. This mechanism will likely be disrupted if alpine glaciers continue to retreat and to disappear [Mark & Seltzer, 2003] with immediate effects on the downstream ecology, which is particularly susceptible to changes in glacier-sourced freshwater input to proglacial streams [Cauvy-Fraunié et al., 2016]. In addition, hydroelectric power production is expected to decrease within the century as a substantial part of the current hydroelectric power is produced by unsustainable glacier mass loss caused by the warming climate [Schaeffli et al., 2019]. As infrastructure grows and glaciers retreat, it will become increasingly important to measure or infer glacier melt runoff, to help to accurately predict its contribution to the catchment's freshwater resources on seasonal and diurnal timescales.

Glacier surface melt is the primary contributor to the mid latitude glacier hydrological system [Shreve, 1972]. However, it remains difficult to observe the dominant processes that drive surface melt with sufficient spatial and temporal resolution [Landmann, 2022]. Conventional *in situ* methods for measuring glacier surface ablation include ablation stakes [Pratap et al., 2015; Fountain & Vecchia, 1999; Landmann et al., 2021] and the use of meteorological data to calculate energy fluxes that result in glacier surface melt [Braithwaite, 1995; Hanna et al., 2005; Lenaerts et al., 2019]. Although ablation stake measurements and reconstruction from meteorological station data are foundational methods, they come with the significant disadvantage of being labor intensive and therefore difficult to implement glacier-wide, long-term studies. Satellite remote sensing, in contrast, offers the only feasible way to monitor glacial melt at a global scale. A wide variety of remote sensing methods have been used to infer glacier surface melt indirectly through observed changes in glacier elevation [Markus et al., 2017; Sutterley et al., 2018], mass [Wouters et al., 2008], or surface

backscatter [Ridley, 1993; Trusel et al., 2013; Bevan et al., 2018]. Although satellite remote sensing may offer true global coverage, it oftentimes lacks the spatial or temporal resolution required to resolve rapid, local variations in surface melt [Yang & Smith, 2013; Yang & Li, 2014; Wille et al., 2019]. More fundamentally, even when remote sensing of glacier surface melt is able to attain a desired spatial and temporal resolution [Trusel et al., 2013; Bevan et al., 2018; Sutterley et al., 2018], such platforms nevertheless benefit from –and in many cases require– *in situ* observations for calibration and validation. Advances in satellite remote sensing of glacier melt therefore motivate the need for improved *in situ* observations of glacier surface melt.

The familiar variety of sounds associated with flowing water attests to the ubiquity of flow-induced acoustics. A correspondingly large number of previous studies have examined the seismo-acoustic wavefield generated by water flow. Basic physical processes implicated in the generation of sound from flowing water include wave breaking [Manasseh et al., 2006], hydraulic jump formation [Ronan et al., 2017], low frequency fluid pulsing in conduits [Podolskiy, 2020], and the entrainment and collapse of air bubbles in turbulent flows [Prosperetti, 1988; Morse et al., 2007]. In terrestrial rivers, both discharge and bedload transport contribute to the seismic wavefield [Burtin et al., 2008, 2011; Gimbert et al., 2016; Roth et al., 2016, 2017; Cook et al., 2018], as do roughness elements such as boulders (and resulting rapids) and engineered blocks and weirs [Schmandt et al., 2013; Osborne et al., 2021, 2022]. In glaciers, flow in subglacial conduits is constrained by conduit size with an observable impact on the seismic wavefield [Bartholomaus et al., 2015; Nanni et al., 2020].

Here, we utilize Distributed Acoustic Sensing (DAS) to record the seismo-acoustic wavefield originating from turbulent supraglacial water flow. The sensing component of DAS is a single mode optical fiber cable deployed on the surface of the glacier. The principle of DAS is that the phase shift of Rayleigh-backscattered light in an optical fiber is used to infer the fiber axial strain rate with spatial resolution on the order of several tens of centimeters and at frequencies, dependent on cable length, of millihertz to several kilohertz [Shatalin et al., 2021], therefore enabling observation of seismo-acoustic wavefields [Lindsey & Martin, 2021; Douglass et al., 2023]. Fluid flow velocities within pipes have been estimated using regression of DAS data [Vahabi et al., 2020; Titov et al., 2022]. Several studies have previously described glacier surface [Walter et al., 2020; Hudson et al., 2021] and borehole DAS deployments [?] for investigating the en- and

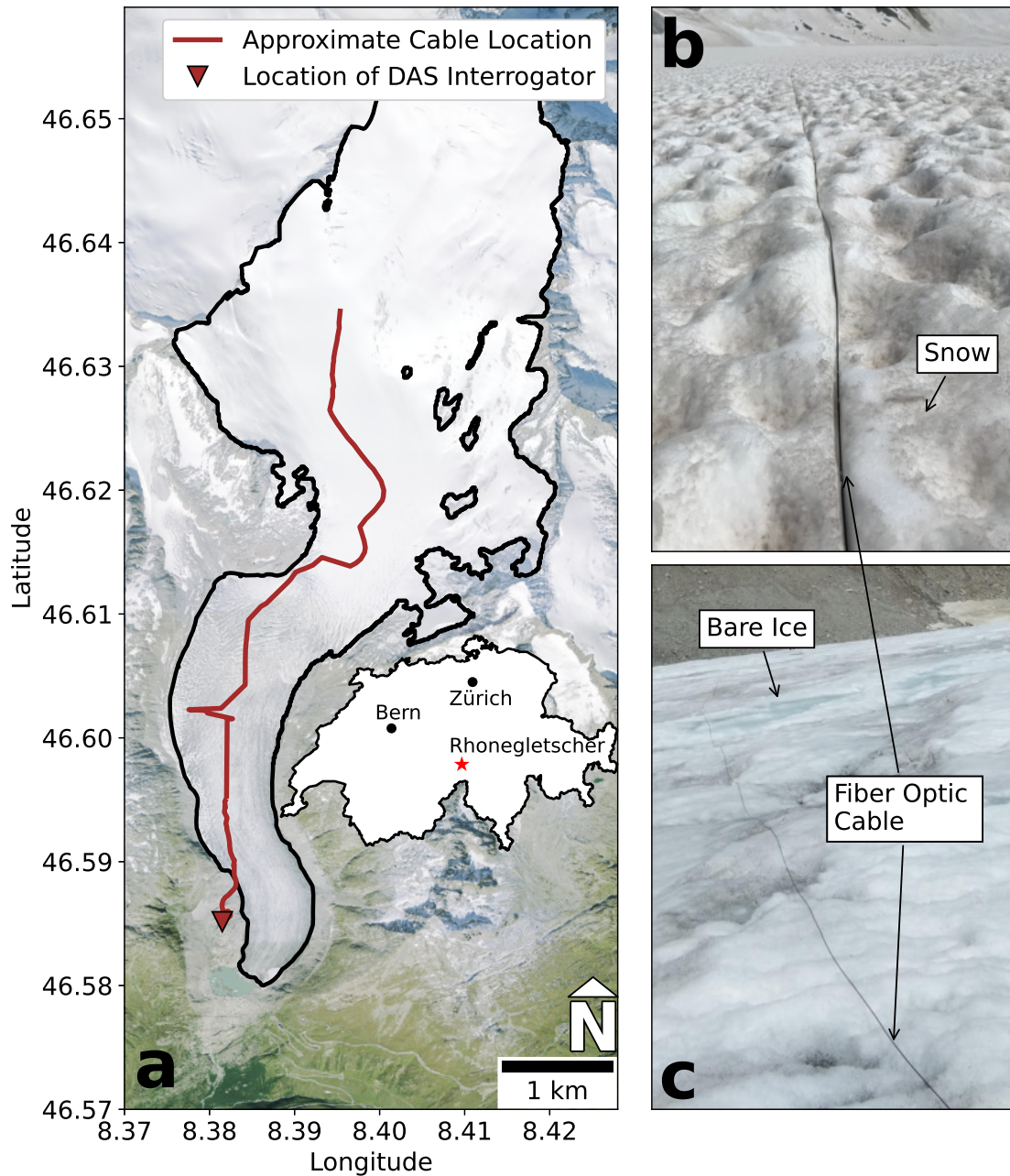


Figure 2.1: a) Map of the study site. Approximate path of the fiber optic cable deployment and location of the Distributed Acoustic Sensing (DAS) interrogator including outline of Rhonegletscher [Consortium, 2005]. Orthophoto provided from the Swiss Federal Office of Topography. b) Photo of the glacier surface and deployed cable in the accumulation zone (credit: Małgorzata Chmiel), consisting mostly of firn at the time of deployment (July 2020). c) Photo of the glacier surface and deployed cable in the ablation zone (credit: Sara Klaasen), consisting primarily of bare ice with areas of crevassing, meltwater surface streams, meltwater pools and glacier moulins.

subglacial environment. Here, we leverage DAS observations from a 9-km long optical fiber deployed along the flow line of an alpine glacier to examine the relationship between glacier melt and the *in situ* glacier surface seismo-acoustic wavefield.

2.3 Field Site and Data

2.3.1 Rhonegletscher

Our measurements were conducted at Rhonegletscher, a temperate mountain glacier located in the central Swiss Alps, in the summer of 2020 (Fig. 2.1a). The glacier covers a total area of 15.5 km² and ranges from 3600 m above sea level (a.s.l.) to 2200 m a.s.l. at its terminus with a length of about 8 km [GLAMOS et al., 2020]. During the field study, the surface of Rhonegletscher in the accumulation zone primarily consisted of firn (Fig. 2.1b). The ablation zone was characterized by bare ice, crevasses, and distributed supraglacial meltwater streams (Fig. 2.1c).

2.3.2 Distributed Acoustic Sensing (DAS) Deployment

A Silixa iDASTM interrogator was deployed in a tent west of the terminus of Rhonegletscher from July 4, 2020 to August 4, 2020. A 9 km single mode fiber optic cable was laid out on the surface of the glacier approximately along the glacier flow line spanning across ablation and accumulation zones. During the first portion of the experiment, interrogator recording settings such as channel spacing and sampling rate were varied for instrument and sensitivity testing. Starting on July 13, settings remained constant for the remainder of the experiment. To avoid complexities with different instrument settings, in this study we only use the data from July 13 to August 4, 2020. During this time data was recorded continuously at 1 kHz sampling frequency, 4 m channel spacing, and 10 m gauge length over 2496 channels. At this sampling frequency, cable length, and gauge length, the iDASTM is sensitive to 2 picostrain per square root Hertz. The last 188 channels contain instrument noise only, because the actual fiber optic cable length was shorter than the length set in the interrogator settings. Thus, we only use the first 2308 channels for our analysis. For most of our analysis, we high-pass filtered the data above 50 Hz. In later analysis, we investigate the unfiltered DAS data to determine the influence of the broad band spectrum on discharge prediction. The

high-pass filter also mitigates the effects of thermal expansion with a diurnal period [Klaasen et al., 2021], shading from transient and local cloud cover, and from other anthropogenic sources [Huynh et al., 2022] such as nearby hydropower production causing narrow-banded seismic energy at 16.7 Hz and 50 Hz. For each channel, we calculated the root mean square (RMS) of the fiber strain-rate for each 30 s window of each channel in the DAS data (Fig. 2.2a).

2.3.3 Discharge measurements

During summer, meltwater from Rhonegletscher is the primary contributor to the highest reaches of the Rhône river near Oberwald, Switzerland. A radar-based discharge gauge (Swiss Federal Office for the Environment, station ID number 2268) located in Gletsch about 3 km downstream of Rhonegletscher’s proglacial lake recorded hourly averaged discharge of the Rhône river throughout the duration of DAS data collection. Discharge data (Fig. 2.2b) were linearly interpolated to 30 s to match the 30-second RMS time steps calculated from the raw DAS data.

2.3.4 Meteorological measurements

We used meteorological data from the station Grimsel Hospiz (Swiss Federal Office of Meteorology and Climatology MeteoSwiss) located 5-8 km southwest of Rhonegletscher behind a mountain ridge. Temperature data were collected at 10 min intervals and precipitation data were recorded as the sum over the 10 min period (Fig. 2.2c).

2.4 Machine learning models

2.4.1 Architectures: Linear, Neural Network, Long Short Term Memory

In order to quantify the relationship between glacier melt and the recorded glacier surface seismo-acoustic wavefield, we employ three separate machine learning models using Keras TensorFlow [Martín Abadi et al., 2015] and assess their relative performance. We first implement a linear model with a single dense layer with linear activation. This model mostly serves as a baseline point of comparison with two more flexible models. Second, we implement a Neural Network (NN) model with two dense layers containing 32 units

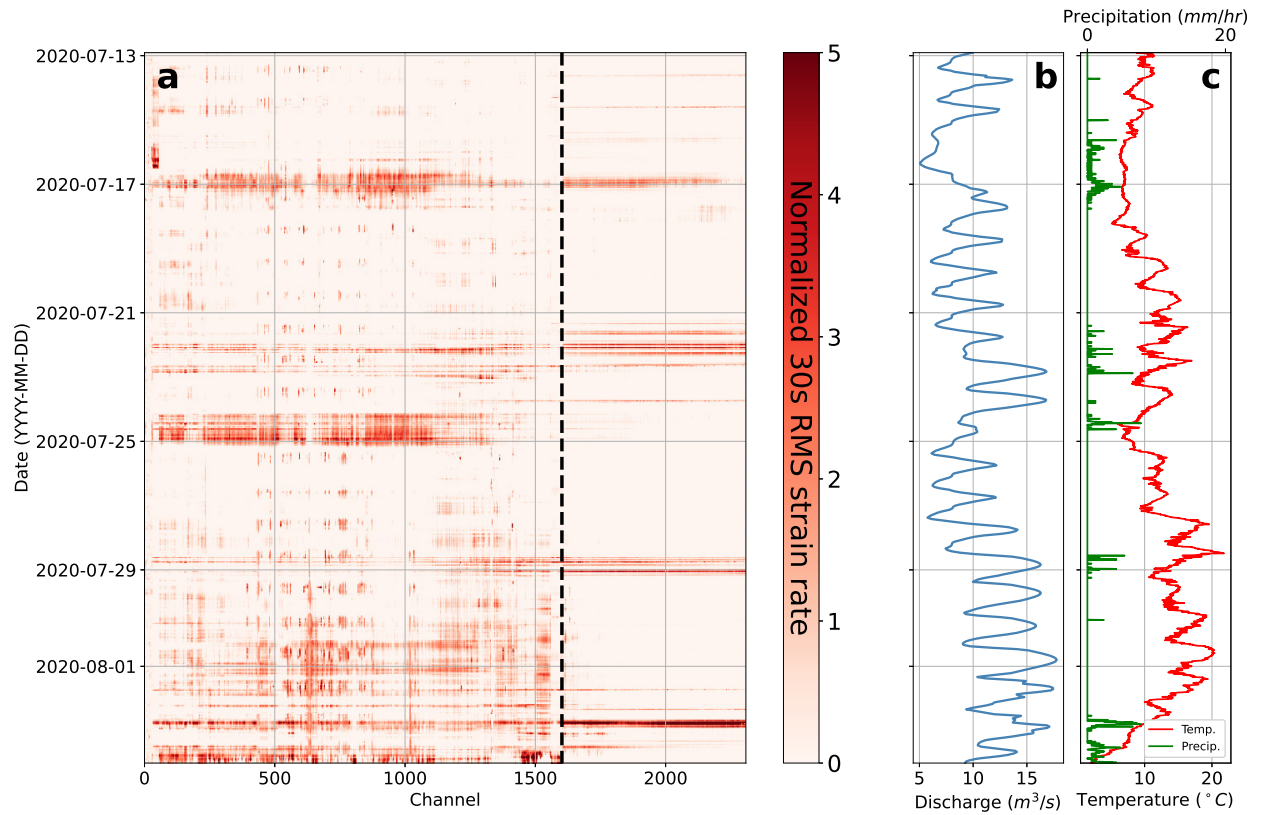


Figure 2.2: **a)** DAS time series over analysis period. Data are highpass filtered above 50 Hz and normalized to peak RMS strain rate over all channels per time step. Low channel numbers are located closest to the terminus down-glacier (i.e. closer to the interrogator) and higher channel numbers are located progressively up glacier according to the plotted cable layout in Figure 2.1a. The dashed line denotes roughly the transition from the ablation zone down glacier and the accumulation zone up glacier. **b)** Rhône river discharge recorded about 3 km downstream of the proglacial lake. During the final two days of the experiment, a standing wave formed in the proglacial stream in the location of the discharge measurement resulting in the three crest pattern that is evident. **c)** Hourly temperature and precipitation data from 10 min recordings at Grimsel Hospitz meteo station (Swiss Federal Office of Meteorology and Climatology MeteoSwiss)

and a rectified linear unit (ReLU) activation function each, a flattening layer, and a dense layer with one unit. Finally, we implement a Long Short-Term Memory (LSTM) model with a single LSTM layer containing 32 units and a dense layer with one unit. The features (independent variables) in our analysis consist of the multivariate time series of DAS strain rate data. The labels (dependent variables) in our analysis consist of the measured discharge values from the downstream discharge gauge. These models are each associated with learning rate, batch size, and data input window size hyperparameters; we choose these hyperparameters based on the results of 90 experiments per model (see Figure A.1). As a result of the analysis, we choose a learning rate of 0.001, a batch size of 32 feature-target pairs, a window size of 200 time steps as these parameters produced stable and robust results. Appendix A further describes hyperparameter tuning.

2.4.2 Cross-validation scheme

Previous studies of changes in supraglacial hydrology through space and time [Nicholson et al., 2021, e.g.] demonstrate that the surfaces of glaciers are inherently non-stationary over the timescale of several weeks during the melt season. Supraglacial stream geometry changes throughout the melt season and responds to change in water flow [Germain & Moorman, 2019]. For this reason, we randomly shuffled the time series windows used for inputs prior to data separation into training, validation, and test sets. We therefore ensure that all possible glacier surface melt regimes occurring during the observation period are captured in the model training data set. In addition to shuffling, we use standard cross-validation (CV) techniques [Bishop & Nasrabadi, 2006, Chapter 14.2] wherein we perform 100 model trainings, each with a uniquely seeded test/training split. CV allows us to quantify model sensitivity to input data and estimate the non-stationary effect of the glacier surface on model performance.

2.4.3 Meteo-LSTM model

We consider an intermediate complexity, “Meteo-LSTM” model that uses an LSTM model architecture with temperature and precipitation data as features and discharge as labels. The goal of this model is to understand the impact of model complexity versus the underlying usefulness of different datasets by testing a model which has similar complexity to the DAS-LSTM model but only relies on the meteo station data.

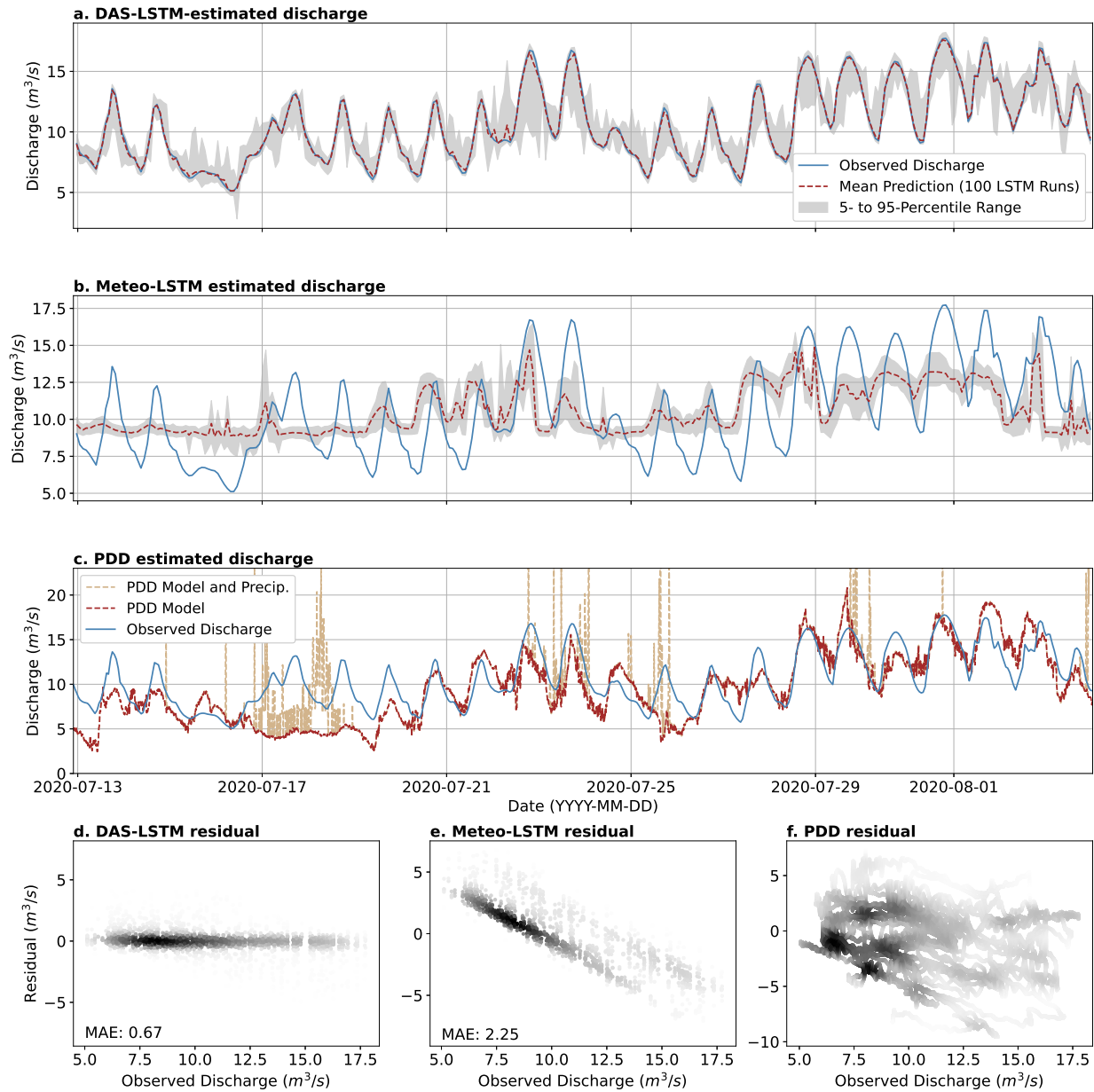


Figure 2.3: a) DAS-LSTM model ensemble mean (red dashed) line and confidence interval (grey region) from cross validation (CV). b) same as a., but with the meteo-LSTM model. c) Positive degree day (PDD) model results. d-f) Residuals for the DAS-LSTM, Meteo-LSTM, and PDD models, respectively.

2.4.4 Positive degree-day (PDD) model

Positive degree-day models are widely used to infer glacier melt from limited meteorological observations [Braithwaite, 1984]. We implement a PDD model following Hock [Hock, 2005]. We carry out a minimization analysis to select the melt rate factor and lapse rate value that resulted in the lowest absolute error in discharge. Temperatures as collected at Grimsel Hospiz were corrected over elevation bands of 100 m. Then the discharge prediction at each elevation band was summed to get the final predicted discharge,

$$D = \sum_{z=2.3 \text{ km}}^{3.6 \text{ km}} \begin{cases} \{[(T + \gamma(z - z_0))] f + P\} A & T > 0 \\ PA & T \leq 0 \end{cases} \quad (2.1)$$

where z is the altitude, z_0 is the terminus altitude, D is the total predicted discharge, T is temperature, γ is the calibrated lapse rate, f is the calibrated melt factor, P is the precipitation rate, and A is the area of the glacier within each step in the summation. The glacier area is given as an idealized rectangle with the glacier area, width, and elevation range as found in GLAMOS [GLAMOS et al., 2020]. The PDD model results were interpolated to match the times of discharge measurements used as LSTM model targets. In order to compensate for meltwater transport from the proglacial lake to the discharge gauge downstream, which is evident from the phase lag between a basic PDD model and measured discharge curves, the PDD model results were shifted based on the phase of maximal cross correlation between modeled and observed discharge.

2.5 Results

The results of our analysis are listed in Table 1. For all of our analyses, we present results in terms of the mean absolute error (MAE) of the residuals and the standard deviation of the residuals between model outputs and discharge gauge measurements. All of these performance statistics are reported for the test dataset in order to quantify model performance when evaluated on data that were not used for parameter estimation. Overall, the best performing models use an LSTM architecture with input DAS data. These models perform about 40% better than the NN model in terms of MAE. The LSTM models also result in a more than 200 times reduction in mean absolute error compared to a linear model.

Table 2.1: Model types and mean absolute error (MAE) for test data set.

Model Type	Input Data	Data Processing	MAE (m ³ /s)	SD (m ³ /s)	Trainable parameters
Linear	DAS	50 Hz High-pass	145.41	232.76	461,601
NN	DAS	50 Hz High-pass	0.88	1.46	81,345
LSTM	DAS	50 Hz High-pass	0.67	1.18	299,681
LSTM	DAS	50 Hz Low-pass	0.68	1.25	299,681
LSTM	DAS	None	0.64	1.15	299,681
LSTM	Meteo	None	2.25	2.74	4,513
PDD	Meteo	None	2.72	3.26	0

We plot the estimated discharge time series and residuals from our model (Figures 2.3a-c and d-f, respectively). Examination of these time series confirms that the DAS-LSTM model is able to capture the phase of discharge (Figure 2.3a). In contrast, the Meteo-LSTM and PDD models, suffer from both poor amplitude and phase response (Figure 2.3b and c).

Model residuals for the DAS-LSTM model show no systematic relationship with increasing discharge (Figure 2.3d). The Meteo-LSTM model, in contrast, show both poor amplitude and phase response (Figure 2.3d) which is likely due to the poor correlation between temperature and precipitation amplitude and phase. The PDD model estimates reasonable amplitudes with a phase shift. We therefore calculate PDD residuals using a best fit time shift. Residuals for the PDD model are uncorrelated with increasing discharge and an order of magnitude larger than the residuals from the DAS-LSTM model.

2.5.1 Ablation zone versus accumulation zone

Models trained on ablation zone data performed better and have less variance than models that only used accumulation zone data. Models trained on data from the ablation zone have a mean MAE of 0.64 m³/s and standard deviation of 0.1 m³/s whereas models trained on accumulation zone data have a mean MAE of 1.07 m³/s and standard deviation of 0.24 m³/s. This can also be seen in the sensitivity analysis discussed in the Discussion section and shown in Figure 2.4 where particular sectors in the ablation zone generally show higher sensitivity to discharge than areas in the accumulation zone.

2.5.2 Meteo-LSTM and PDD results

The results of LSTM model runs with temperature and precipitation as inputs are shown in Figure 2.3b. The PDD predictions were shifted according to highest correlation coefficient, corresponding to 5.4 hours, before the residual was calculated to account for meltwater transport to the discharge gauge. The MAE of the residuals of the predictions on the test sets of data are $2.29 \text{ m}^3/\text{s}$ and $2.27 \text{ m}^3/\text{s}$ for the Meteo-LSTM and the PDD models, respectively.

2.5.3 Low-frequency versus high-frequency

Models trained on low frequency ($<50 \text{ Hz}$) filtered DAS data perform slightly worse with a MAE of $0.68 \text{ m}^3/\text{s}$ compared to $0.67 \text{ m}^3/\text{s}$ of the high frequency trained models while also having a larger residual standard deviation of $1.25 \text{ m}^3/\text{s}$ compared to $1.18 \text{ m}^3/\text{s}$ of highpass filtered models. An analysis of 100 LSTM models trained on unfiltered DAS data was also done which performs slightly better than both filtering methods with an MAE and standard deviation of $0.64 \text{ m}^3/\text{s}$ and $1.15 \text{ m}^3/\text{s}$ respectively which may be explained by the broadband nature of the surficial hydrological soundscape [Podolskiy et al., 2023].

2.6 Discussion

Our study demonstrates the potential for DAS-based glacio-hydrological sensing to be a robust technique for potential in situ measurements of glacier runoff. We find good agreement with $0.64 \text{ m}^3/\text{s}$ MAE between DAS-LSTM-inferred and stream gauge-measured discharge values. We begin this section by discussing why the seismo-acoustic wavefield carries so much correlation with glacier discharge.

2.6.1 The physical basis relating discharge to the seismo-acoustic wavefield

As described in the Introduction, a wide variety of processes contribute to the glacier seismo-acoustic wavefield. A key result that allows us to decipher the origin of our wavefield–discharge relationship is that our regression analysis performs equally well or slightly better in the range 50-500 Hz as compared to the range 0-50 Hz. This high frequency band eliminates the possibility that the dominant signal in our analysis has its origin in subglacial processes such as conduit flow [Bartholomaus et al., 2015], gurgling crevasses [Podol-

skiy, 2020], and bedload transport [Roth et al., 2016, 2017], all of which are thought to create signal below 50 Hz. Furthermore, crevassing and basal stick-slip sliding is expected to generate seismic signals above 50 Hz [Podolskiy & Walter, 2016] in addition to anthropogenic activity and wind [Podolskiy et al., 2023] will also cause increased RMS. However, we infer that the sound generated from supraglacial streams is the dominant contributor to our discharge regression analysis due to its persistent existence during our melt-season measurement. Our basis for this inference is by comparison with previous studies that have examined the same acoustic frequency range in the context of terrestrial rivers [Bolghasi et al., 2017; Osborne et al., 2021, 2022; Podolskiy et al., 2023]. Additionally, Figure B.3 compares wind from the nearby meteo station to daily means of DAS strain rate and variance RMS observations and show little correlation throughout the experiment which suggests that supraglacial turbulent flow to be the dominant signal. However, we emphasize that supraglacial turbulent flow is not the sole contributor to the seismo-acoustic wavefield and on glaciers with less prominent supraglacial runoff, it may become more important to disentangle seismic signals generated from other processes from the signal generated from the supraglacial hydrological system.

2.6.2 DAS offers a stable observation platform on melting glacier surfaces

Ensuring the stability of instrumentation on the surface of glaciers is notoriously challenging [Carmichael, 2019]. As a result, most melt season seismic deployments in the ablation zone of glaciers, for example, only cover spatial apertures on the order of 1 km [e.g. Rösli et al., 2014]. Studies that have employed dense glacier surface arrays have generally avoided the melt season due to melt-induced tilt and toppling of the instruments [Gimbert et al., 2021]. Stream discharge in terrestrial rivers is usually measured by establishing a relationship, called a rating curve, that empirically relates stream height (also called stage) to discharge [Kennedy, 1984]. In order to bypass logistical complexities associated with this approach, recent studies have elected to pursue passive acoustic observation of river height [Osborne et al., 2021; Podolskiy et al., 2023]. The motivation to use seismo-acoustic observations to study surficial glacier hydrology is even stronger given that seasonal variations in stream morphology [Knighton, 1981; Marston, 1983; Karlstrom et al., 2013] would be expected to result in a strongly time-dependent rating curve. For our study, the deployment of the cable along the glacier flow line allows for sensitivity to source mechanisms in a wide area encompassing both the ablation and accumulation zones. A particular benefit of fiber optic sensing

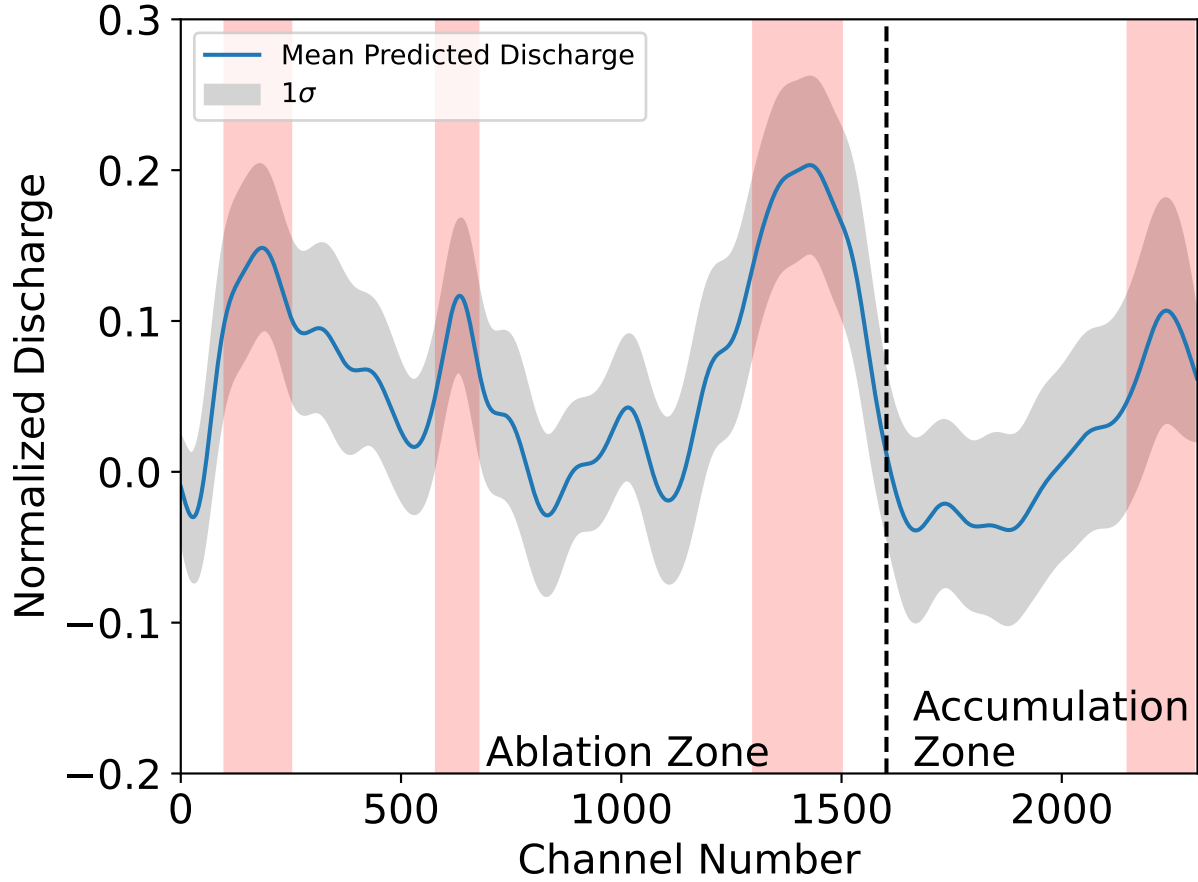


Figure 2.4: Channel sensitivity analysis from applying a uniform in time Gaussian pulse with a width of 50 channels. A new discharge prediction is made each time the Gaussian pulse is centered on the next channel. The mean prediction is calculated from the predicted discharge of the 100 LSTM models produced. Predictions are given in values of a normalized discharge. A spatial trend in discharge sensitivity arises at four locations highlighted in red: three sectors in the ablation zone and one sector in the accumulation zone. At these locations, a given increase in normalized strain rate results in higher predicted normalized discharge values than would be expected at other locations along the cable. The dashed line denotes the approximate location of the transition from the ablation zone to the accumulation zone as determined by the drop in correlation of strain rate RMS with wind speed which reflects the cable melting into snow. This point had moved roughly a kilometer up glacier over the course of the experiment and may explain the significant peak in predicted discharge near the transition line.

over other methods is that the fiber optic cable can be deployed strategically and is not limited to the specific instrumentation requirements such as the availability of electrical power at the sensing location that hinder many other types of seismic sensing equipment or other in situ instrumentation. In this study, the cable transects many features typical of mountain glaciers: crevasses, supraglacial streams, rock debris, firn, snow, etc.

2.6.3 Sensitivity Analysis

All model iterations show a spatial sensitivity to predicting discharge. In Figure A.2, we investigated prediction performance relative to different parts of the cable by isolating the observed acoustic noise in these locations. The DAS data was sectioned in three different ways and used as model input to predict discharge: the whole cable, only channels within the ablation zone, and only channels within the accumulation zone. We find model improvement when data within the ablation zone, where we expect the most pervasive surface hydrology to exist, is used for training and prediction. When only the data from the accumulation zone is used for training and prediction, the models perform markedly worse, $1.03 \text{ m}^3/\text{s}$ mean MAE as compared to $0.63 \text{ m}^3/\text{s}$ mean MAE for the ablation data. In addition, the standard deviation of the residuals is three times higher than that of the models using ablation data alone. In the following subsections, we discuss possible mechanisms by which changes in the meltwater flow within supraglacial streams as a result of temporal variation in discharge cause fluctuations in acoustic noise power as observed by DAS.

Figure 2.4 shows a model sensitivity analysis where we generate a synthetic strain rate Gaussian pulse with a width of 50 channels and uniform in time. The pulse is then centered on each channel before making a discharge prediction. We iterate this procedure for each LSTM model trained on the whole cable DAS data. Increased values of predicted normalized discharge for a given channel in Figure 2.4 indicate that an increase of measured DAS strain rate or acoustic noise results in an increase of predicted discharge. Three sectors of cable in the ablation zone centered around channel 150, 650, and 1400 are shown to be of more importance to predicting discharge from DAS strain rate. Interestingly, a sector around channel 2250 in the accumulation zone near the glacier headwall also imparts some sensitivity to predicted discharge. The most sensitive portion of the cable is the sector around channel 1400 where the snow line is located during the cable deployment time and the ice fall of Rhonegletscher is located. The melting of snow around the snow

line during the observation period caused the snow line to recede and exposed more bare ice to the fiber. Surface crevassing, newly formed meltwater streams, and audible drainage from within exposed crevasses may have all contributed to the high RMS strain rate signal in this area. This provides a first step into the potential of forming a spatially-distributed, rather than integrated, inference of glacier surface melt.

2.6.4 Scaling up to other glaciers, longer time spans, and the potential for monitoring

We have demonstrated that DAS can be used to infer glacier runoff on Rhonegletscher. The model that we have trained is not expected to be immediately portable to other glaciers, however, for the simple reason that changing the layout of the cable would result in different channel weightings. Glaciers that have differing contributions of runoff and glacier melt to total discharge may require further independent discharge measurements, at least initially, to validate the model inference. Despite these complications, the acoustic noise-discharge relationship does appear to persist with a variety of flow regimes [Podolskiy et al., 2023] which we expect to be the case for supraglacial DAS observations as well.

We have demonstrated discharge inference over a one month time period. We do not expect our model to perform well when extrapolated to more diverse glacier surface conditions than those encountered during our deployment. Additional observations would likely be necessary to capture the wide variety of surface energy balance regimes that occur over long time periods. We expect this to be true, for example, when comparing end-member summertime and wintertime conditions [e.g., Chapter. 5, Cuffey & Paterson, 2010]. However, it may also be the case that smaller term variations, for example due to supraglacial stream rearrangement during the summer melt season [Pitcher & Smith, 2019], may also require more detailed training data in order to match the performance attained by our model. Over these long time periods, a number of environmental factors could change and thereby limit the performance of our model. These factors include the coupling of the fiber to the surface of the glacier and, at longer time scales, even the geometry of the cable. During our deployment, the coupling of the fiber in certain areas did change (e.g. melting of snow around the snow line), the effect of which is accounted for in our cross-validation scheme and shows no significant reduction in performance through time (see Figure 2.3a). For this reason, we suggest that minimal model retraining would be required on account of changes in fiber coupling alone.

Our data analysis was retrospective, however, there is no fundamental reason why a trained model such

as ours could not be used for near-real time discharge estimation. Once a model is trained on DAS data and tested for accuracy, it can be applied to previously unseen DAS data recorded from the same fiber array to infer discharge. Carrying out prediction with a our trained model is orders of magnitude faster than the training and testing steps. Depending on the location of a discharge gauge downstream, this time limitation may be sufficiently faster than a reading from a proglacial discharge gauge and indicates an ability to supplement traditional discharge monitoring networks.

Our approach could provide particularly useful information in several glaciological settings. The most salient example is for glaciers that terminate in the ocean or in lakes. For these glaciers, it is not possible to deploy traditional stream gauges and discharge is generally estimated through melt stakes or surface energy balance calculations [Jackson et al., 2022]. In glaciated catchments that have complex networks of proglacial streams or multiple points of runoff, traditional discharge instrumentation would become logistically burdensome.

It is not known at the present time whether it would be possible to train a model on sufficiently many different glaciers or cable layouts so as to arrive at a general model that would capture the relationship between DAS and discharge for an arbitrary new glacier where discharge measurements have yet to be made.

2.7 Conclusion

In situ measurements of glacier runoff have previously been logistically difficult to obtain, particularly in areas with geographically complicated catchments or glaciers with distributed surface hydrological regimes. We demonstrate a correlation between the *in situ* seismo-acoustic wavefield measured from the surface of a glacier and proglacial discharge measured by a radar-based gauge. Our machine learning model that relates these quantities identifies spatial variability and coherence in discharge sensitivity to acoustics. The ability to quantify glacier runoff using turbulent flow generated seismo-acoustics as observed by DAS opens the door to gaining insights into these regions. Discharge inferences produced by DAS and ML could one day be ingested in glacier mass balance models that have typically been limited by a lack of *in situ* glacier runoff validation [Lenaerts et al., 2019]. In addition, seasonality of accumulation, ablation, and runoff may be characterized by changes in acoustic signals that we observe here during the melt season; however, this will

need to be investigated in subsequent studies. Here, we have demonstrated the first of its kind application of DAS for inferring glacier runoff informed by radar-based discharge and other observations.

Chapter 3

Ice flow controls on old ice preservation in the Allan Hills, Antarctica

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3.1 Chapter Abstract

Ice cores recovered from the Allan Hills Blue Ice Area, Antarctica, contain the oldest ice yet found on Earth. Their exceptional age provides direct constraints on atmospheric composition and climate during intervals associated with major reorganizations of Earth's climate system. However, the flow mechanisms responsible for preserving old ice are not fully understood. Here, we develop a three-dimensional thermomechanical ice-flow model to characterize regional present-day flow, and interpret the results in the context of surface velocity measurements, borehole temperature profiles, and ice-core geochemistry. Our model reveals nearly stagnant basal ice upstream of a local nunatak, where the oldest ice has been recovered, linking older ice with slower modeled basal flow. Conversely, the nearby Cul-de-sac zone preserves old ice despite faster modeled flow. Distinct Cul-de-sac geochemical signatures indicate a different source region, consistent with transport from elsewhere rather than local preservation by stagnation. These results show that old ice is preserved through at least two endmember modes: local preservation in zones of stagnant flow, and imported preservation through advection of old ice from upstream. These results provide a framework for

understanding the dynamic pathways of old ice preservation and help guide future efforts to locate and interpret Earth’s oldest ice.

3.2 Introduction

Discontinuous ice cores drilled in the Allan Hills blue ice area (BIA) contain the oldest ice yet recovered on Earth, reaching ages of ~ 1.2 Ma [Higgins et al., 2015], ~ 2.7 Ma [Yan et al., 2019], and >6 Ma [Shackleton et al., 2025]. These records have the potential to provide unique constraints on atmospheric composition and climate during key intervals of Earth’s climate evolution [Marks-Peterson et al., 2026; Shackleton et al., 2026]. However, the Allan Hills cores are stratigraphically complex [Higgins et al., 2015; Yan et al., 2019; Shackleton et al., 2025; Kirkpatrick et al., 2025; Stoll et al., 2026], reflecting a complex ice flow history. Understanding the mechanisms that preserve and expose old ice at the surface is therefore essential for interpreting these records and for identifying future drilling targets. Antarctic BIAs typically occur in regions of persistent ablation near nunataks or steep bed topography [Bintanja, 1999; Bintanja & Reijmer, 2001; Tollenaar et al., 2024; Hui et al., 2014]. In these environments, strong katabatic winds remove snow and promote sublimation, allowing older ice that has been transported into the ablation zone to be exposed at the surface [Bintanja, 1999]. The terrestrial age of meteorites transported into Antarctic BIAs record the age of the ice that transported them to the surface and can be used as an indicator for the presence of old ice [Scherer et al., 1997]. Previous work has suggested that interactions between ice flow and bedrock topography play a central role in shaping BIA dynamics. Simplified ice-flow models have been used to explain meteorite concentration [Benoit, 1995] and to investigate isochrone structure within BIAs [Kehrl et al., 2018; Grinsted et al., 2003], while laboratory experiments demonstrate that ice can slow and stagnate upstream of bedrock rises [Corti et al., 2008]. More recently, three-dimensional (3D) modeling studies have explored the influence of ice rheology and anisotropy on flow in Antarctic BIAs [Zwinger et al., 2014]. Despite these advances, quantitative descriptions of ice flow in regions of multimillion-year-old ice that can be directly compared with modern geophysical observations remain limited.

Here we combine 3D ice-flow modeling with geophysical and geochemical observations to investigate the mechanisms responsible for the preservation of old ice in the Allan Hills BIA. We develop a 3D thermo-mechanically coupled full-Stokes model using high-resolution bed topography derived from ground-based

radar surveys [Nesbitt & Brook, 2023], together with BedMachine [Morlighem et al., 2017] and the Reference Elevation Model of Antarctica [Howat et al., 2022]. Full-Stokes models consider the full 3D stress balance terms and are useful in reproducing observed velocities in areas of large ice thickness gradients and steep bed topography, where simpler models fail [Le Meur et al., 2004]. Model predictions are evaluated using surface velocities measured with a network of Global Navigation Satellite System (GNSS) stakes deployed during the NSF-COLDEX 2022–23 and 2023–24 field campaigns, as well as borehole temperature profiles that constrain vertical ice motion through thermal advection, also measured during COLDEX campaigns.

Our results reveal a zone of slow to nearly stagnant [Hiemenz, 1911; Batchelor, 1967] ice upstream of the nunatak near the ALHIC1902 and ALHIC2302 boreholes (see Fig. 3.1), where the oldest ice in the region has been recovered. At these boreholes, the maximum age of the ice correlates with slow modeled basal flow, consistent with local preservation in zones of stagnation. However, nearby sites within the Cul-de-sac (CDS), a pronounced local surface depression downstream of a significant surface gradient, preserve old ice up to 5.32 Ma despite substantially faster modeled flow. Distinct $\delta^{38}\text{Ar}/^{36}\text{Ar}$ and $\delta^{18}\text{O}$ signatures indicate that this ice originated in a different accumulation environment. Together, these observations show that old ice exposures in the Allan Hills arise through two distinct dynamic pathways: local preservation in zones of slow or stagnant flow and advection of ancient ice from a distinct upstream source region.

3.3 Main

The modeled and observed ice-flow field in the Allan Hills BIA is strongly three-dimensional and reflects the interaction between regional ice-sheet flow and local topography. Regional ice flow enters the model domain from the grid northeast, consistent with large-scale ice-sheet flow toward the Allan Hills region. As this flow approaches the nunatak near the ALHIC1902 and ALHIC2302 boreholes, it is diverted laterally around the exposed bedrock (Fig. 3.1).

A locally sourced downslope component originating from the flanks of the nunatak is superimposed on this regional flow. Ice flowing down the nunatak moves in the opposite direction of the incoming regional flow (Fig. 3.1b). The interaction between these two flow components produces a localized zone of convergence on the upstream side of the nunatak. Convergence is accommodated through three-dimensional flow

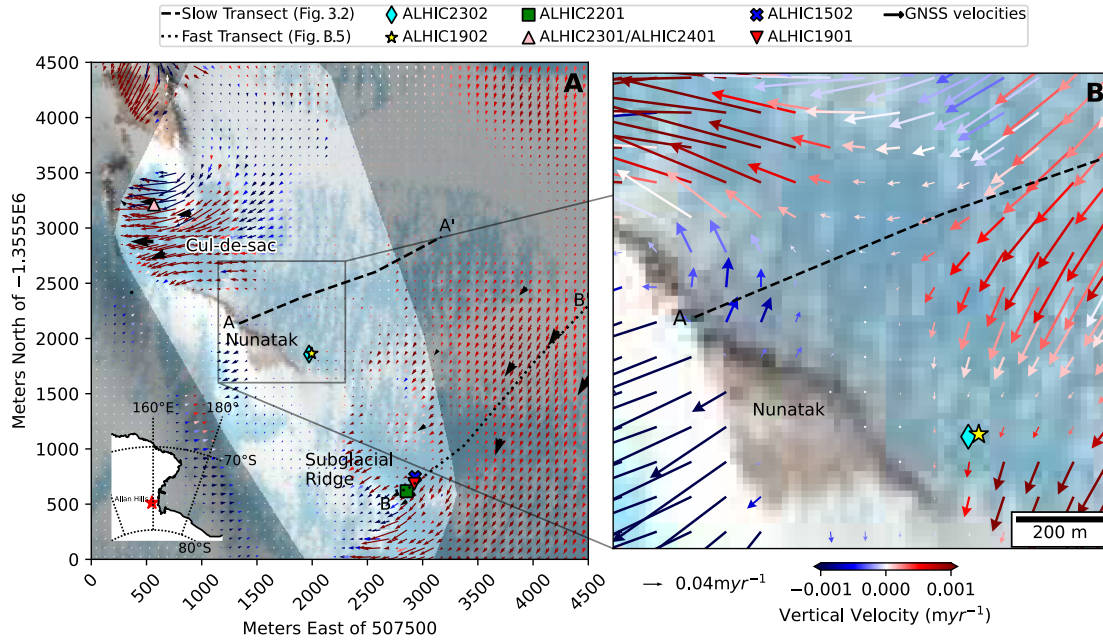


Figure 3.1: Modeled ice flow 10 m above the bed. **a)** Ice-flow field across the fine-mesh model domain. The inset shows the location of the Allan Hills in Victoria Land, Antarctica. The highlighted region marks the extent of ground-based ice-penetrating radar data collected during the 2019–20 field campaign [Nesbitt & Brook, 2023], while the shaded region indicates areas where BedMachine [Morlighem et al., 2017] provides bed elevation in the model domain. **b)** Zoomed view of the upstream side of the nunatak showing convergence between regional ice flow and locally sourced downslope flow from the nunatak, producing a zone of slow or nearly stagnant ice. Velocity vectors are exaggerated to emphasize flow direction. Dashed lines indicate transects through the area of stagnation (black, Fig. 3.2) and faster ice flow (gray, Fig. B.5) to show the englacial flow field.

around the nunatak and through a pronounced reduction in basal ice velocity near the base of the bedrock rise.

Our 3D thermo-mechanically coupled ice-flow model quantifies this flow structure and reproduces the GNSS observed (see Table B.1) deflection of ice around the nunatak (Fig. B.1). The model predicts a zone of very slow ice motion immediately upstream of the nunatak where the opposing flow components interact. This region coincides with the locations where the oldest ice cores in the Allan Hills BIA have been recovered.

3.3.1 Basal stagnation near the nunatak

Modeled basal ice velocities vary strongly across the study region. Along the grid eastern flanks of the nunatak, a downslope counterflow toward the ice sheet reaches horizontal velocities as high as 20 mm yr^{-1} with vertical velocities -1 mm yr^{-1} (downwards). The resulting reduction in basal ice motion produces a region of nearly stagnant ice immediately upstream of the nunatak, near the bed of the ice sheet. The average velocity of the bottom 20 m of ice are 0.5 mm yr^{-1} and 0.4 mm yr^{-1} at ALHIC1902 and ALHIC2302, respectively. Averaged over the entire ice column from the surface to bedrock, horizontal and vertical velocities at these boreholes are 4 mm yr^{-1} and 5 mm yr^{-1} .

Stagnation is absent in the faster-flowing region grid southeast of the nunatak, where ice flows across the southward extension of the nunatak as it dives beneath the ice surface and continues as a subglacial ridge. The corresponding basal ice velocities in this region are 12 mm yr^{-1} and 9.5 mm yr^{-1} at ALHIC1901 and ALHIC1502, respectively. In this faster flowing ice, the horizontal and vertical column average velocities are 37 mm yr^{-1} and 21 mm yr^{-1} , respectively. Even faster basal motion is modeled in the CDS zone (boreholes ALHIC2301 and ALHIC2401), where basal ice velocities are 46 mm yr^{-1} and 49 mm yr^{-1} , respectively. In the CDS, modeled horizontal velocities reach 48 mm yr^{-1} and vertical velocities approach 8 mm yr^{-1} due to strong ice thickness gradients.

In Figure 3.2, we present ice velocity along a vertical cross-section. A large portion of the basal ice, about 20 m, flows slower than $\sim 0.5 \text{ mm yr}^{-1}$. At this speed, the ice would only flow 500 m in 1 Myr and could remain within the area of convergence and stagnation. Even if basal ice accelerated to the depth-averaged velocity of $\sim 2 \text{ mm yr}^{-1}$ —for example during a previous glacial period when a thicker ice sheet

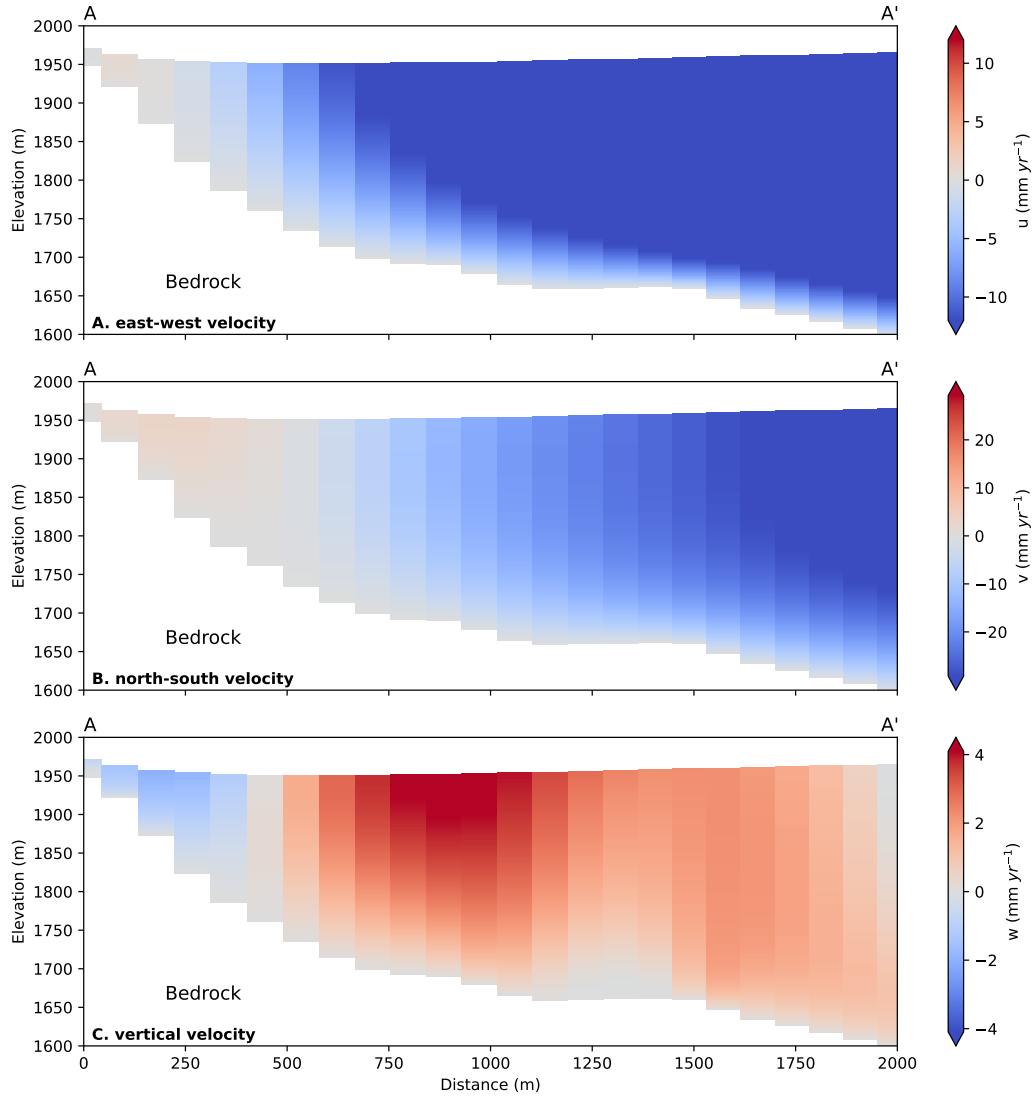


Figure 3.2: Cross-sectional profiles of modeled ice velocity along the transect shown in Fig. 3.1a in the **a)** grid north–south (u component of velocity), **b)** grid east–west (v component of velocity), and **c)** vertical (w component of velocity) directions. In **a)** positive is in the grid west direction and negative is in the grid east direction; in **b)** positive is in the grid north direction and negative is in the grid south direction; and in **c)**, positive is in the upwards direction and negative is in the downwards direction. Downslope flow originating from the ALHIC1902/2302 nunatak opposes the incoming regional ice-sheet flow, producing convergence and a zone of slow or nearly stagnant basal ice near the base of the nunatak.

overrode the presently exposed nunataks [Atkins et al., 2002]—ice would remain within a ~ 500 m long stagnation zone for ~ 250 kyr. Even this lower-bound for the residence time suggests that ice resides within the stagnation zone for multiple glacial–interglacial cycles.

3.3.2 Constraints on ice flow from borehole temperature observations

Borehole temperature profiles provide an independent constraint on ice motion. In polar ice, ice temperature reflects the combined effects of thermal conduction, geothermal heating, and advective heat transport by ice flow [Hills et al., 2017]. Deviations from a purely conductive temperature profile therefore record the integrated influence of ice motion and past climate. Because ice rheology is strongly temperature dependent, measured temperatures at depth also provide an important validation for the thermo-mechanical ice-flow model used in this study (see Fig. B.2).

We find that temperature measurements from Allan Hills boreholes cannot be explained by a steady-state conductive–geothermal balance alone. Modeled temperature profiles that include advective heat transport and climate warming signal more closely reproduce the observed borehole structure (Fig. B.2). When comparing observed borehole temperatures to a temperature simulation that does not include advection, modeled temperature profiles of boreholes located in regions of relatively fast ice flow (e.g., ~ 10 mm yr $^{-1}$) have a greater mean absolute error (MAE) from observed borehole temperature profiles than boreholes in regions of predicted slow or nearly stagnant flow (e.g., ~ 0.1 mm yr $^{-1}$) (Fig. B.2m).

Because borehole temperature inversions are non-unique and sensitive to past climate history [Clow, 1992], we include ~ 2 K of surface warming over the past ~ 150 years in our simulations, consistent with regional estimates [Yang et al., 2018]. Even with this imposed warming history, the observed borehole temperature profiles require spatially variable advective heat transport, consistent with the heterogeneous velocity field predicted by the ice-flow model (Fig. B.2). Together with the GNSS velocity measurements, these observations confirm that the model captures the first-order structure of the present-day velocity field.

3.3.3 Relationship between basal velocity and ice age

We next examine the relationship between the present-day modeled flow field 10 m above the bed and the maximum age of ice preserved at each drill site (Fig. 3.3a). Allan Hills BIA ice upstream of the nunatak and

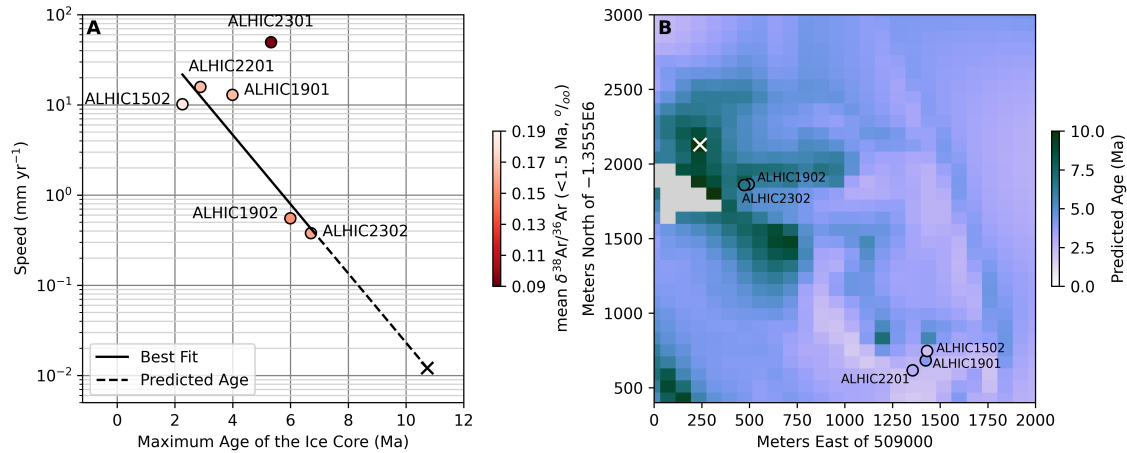


Figure 3.3: **a)** Relationship between modeled basal ice flow speed or the magnitude of the components of flow (10 m above the bed) and the maximum ^{40}Ar dated age recovered from each ice core. Ice cores are colored by the mean $\delta^{38}\text{Ar}/^{36}\text{Ar}$ of samples younger than 1.5 Ma. For sites outside the Cul-de-sac (CDS), maximum age increases as modeled basal flow decreases, indicating local preservation of old ice in zones of slow flow. The solid line shows a best-fit trend for these sites and the dashed line shows an extrapolated age according to the best-fit line found from observations. The X marks the oldest predicted ice, age 10.73 Ma, in the region of interest. The CDS site (ALHIC2301) deviates from this relationship and is isotopically distinct, suggesting preservation of old ice transported from a different source region. **b)** Predicted maximum ice age distribution near the nunatak and subglacial ridge using the relationship shown in **a)** by the dashed line of best-fit. The color of scatter points in **b)** refer to the measured maximum age of the ice core. The X marks the location of the oldest predicted ice age as found in **a)**.

subglacial ridge shows a clear relationship between ice age and basal ice flow speed. Ice cores with younger maximum ice ages are associated with higher modeled basal ice flow speeds, whereas ice cores preserving older ice correspond to slower modeled flow near the bed. This trend suggests that extremely slow basal flow promotes the local preservation of old ice by allowing ice parcels to remain within a zone of flow stagnation for extended periods of time. The predicted distribution of old ice in the Allan Hills BIA using this ice age-flow speed relationship is shown in Figure 3.3b and reveals a broad area of old ice upstream of the nunatak with pockets of old ice where ice flow intermittently slows fitting well with the observed ice core age records. This relationship supports the calculations of ice transit time through areas of stagnation made above.

3.3.4 The Cul-de-sac as a distinct flow regime

One site deviates strongly from the velocity-age relationship described above. ALHIC2301, located in the Cul-de-sac (CDS), preserves very old ice despite comparatively rapid modeled basal flow and measured surface velocity. Modeled horizontal velocities near ALHIC2301 are nearly forty times larger than those at the boreholes experiencing stagnation, reflecting the strong surface slope and associated ice-thickness gradients driving flow into the CDS.

Modeled velocity fields and GNSS observations indicate that ice enters the CDS primarily from the grid east-northeast (Fig. 3.1). Unlike the stagnation boreholes upstream of the nunatak, where slow basal motion allows ice to remain locally preserved, the CDS appears to receive ice advected from upstream regions.

Geochemical measurements from the ice cores support this interpretation. To further assess ice provenance, we examine $\delta^{38}\text{Ar}/^{36}\text{Ar}$ ratios, which are related to the firn thickness of the ice core accumulation zone [Severinghaus et al., 2003] as well as $\delta^{18}\text{O}$ of the ice, a proxy for the temperature of the site of snow accumulation [Masson-Delmotte et al., 2008]. We limit our geochemical analysis to ice from each ice core dated <1.5 Ma because $\delta^{18}\text{O}$ shows no spatial distinction in ice >1.5 Ma and $\delta^{38}\text{Ar}/^{36}\text{Ar}$ values are altered in older samples near the bed [Shackleton et al., 2025]. We find that for the youngest (<1.5 Ma ice), the CDS core (ALHIC2301) is distinct from the other cores. While the ALHIC2301 ice core has a mean $\delta^{38}\text{Ar}/^{36}\text{Ar} = 0.094\text{‰}$ (standard error, $\text{SE} = 0.007\text{‰}$), the mean $\delta^{38}\text{Ar}/^{36}\text{Ar}$ for the cores outside of the CDS range from $0.153 - 0.183\text{‰}$ ($\text{SE} = 0.002 - 0.012\text{‰}$) (Fig. 3.3). A similar pattern is seen in $\delta^{18}\text{O}$ of the ice (Fig. B.3): the CDS ice core shows significantly lower mean $\delta^{18}\text{O} = -39.75\text{‰}$ ($\text{SE} = 0.384\text{‰}$) compared to a range of mean $\delta^{18}\text{O} = -37.58$ to -36.93‰ ($\text{SE} = 0.072 - 0.337\text{‰}$) in the other ice cores. Together, these observations indicate that the CDS differs dynamically from the stagnation boreholes upstream of the nunatak.

3.4 Discussion and conclusions

Our results provide a new framework for understanding how multimillion-year-old ice can be preserved and exposed within Antarctic BIAs. By combining a high-resolution 3D thermomechanical ice-flow model with surface velocity measurements, borehole temperature profiles, and ice-core geochemistry, we show that old ice in the Allan Hills BIA is preserved through two distinct dynamical pathways. These pathways

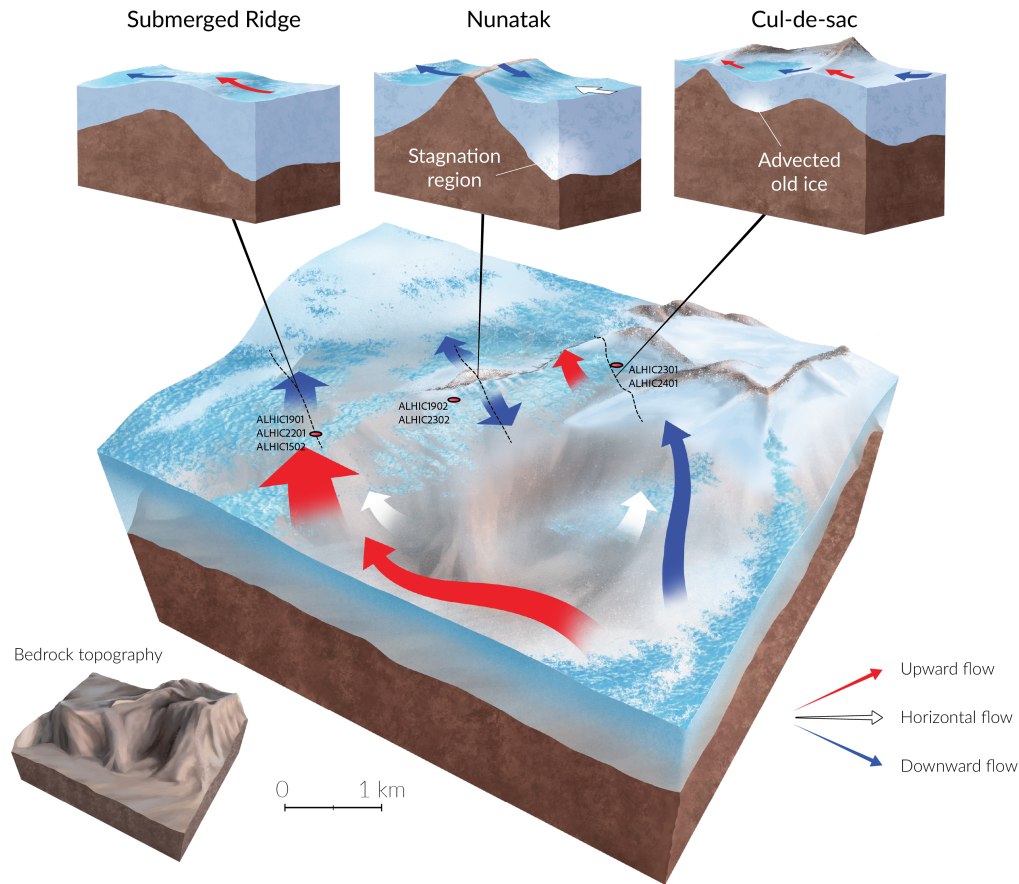


Figure 3.4: Conceptual overview of ice flow in the Allan Hills, Antarctica. Three distinct flow regimes are observed over the submerged ridge, on the flanks of an exposed nunatak, and a region of complex topography called the Cul-de-sac (CDS).

correspond to (1) local preservation within zones of extremely slow or stagnant basal flow, and (2) imported preservation of old ice from an upstream source region. This interpretation reconciles the spatial patterns of ice-core age, modeled ice velocities, and isotopic signatures observed across the study area. A conceptual overview of our results is shown in Figure 3.4.

A striking aspect of these results is that the modern flow field encodes the long-term structure of ancient ice preservation. Our results show that the locations of the multimillion year ice are interpretable in the context of the present-day flow configuration, despite ice ages into the late Miocene (5.3 Ma, see Fig. 3.3). Over such timescales, the geometry and thickness of the East Antarctic Ice Sheet have likely varied through its initiation and subsequent variation within glacial–interglacial cycles [Atkins et al., 2002; Delisle

& Sievers, 1991]. Although regional geological records from the Transantarctic Mountains provide important constraints on long-term ice-sheet variability, direct constraints on ice thickness at the spatial scale of the Allan Hills region remain sparse. Nevertheless, the modern flow field captures the dynamical conditions that determine where ancient ice is preserved today.

Locally generated downslope flow from a nunatak opposes regional ice sheet flow and produces zones of convergence and stagnation. Previous studies of Antarctic BIAs generally emphasize upward ice flow toward the Transantarctic Mountains driven by large-scale ice-sheet geometry [Whillans & Cassidy, 1983; Jr et al., 1983; Sinisalo & Moore, 2010], a pattern that is more applicable to the CDS where ice cannot readily escape. In contrast, our results show that ice descending from the nunatak can oppose this regional flow, creating the stagnation zone upstream of the ALHIC1902 and ALHIC2302 boreholes. A similar interpretation of convergent flow upstream of a bedrock rise was proposed by Coren et al. [Coren et al., 2003], although they analyzed topography that remained fully submerged beneath the ice surface rather than forming an exposed surface feature.

3.4.1 Local preservation by stagnation

The oldest ice recovered near the ALHIC1902 and ALHIC2302 boreholes is associated with a region of extremely slow basal motion produced by the interaction of regional ice-sheet flow with locally sourced downslope flow from the nunatak. Our model shows that ice flowing from the interior (grid northeast) is diverted around the nunatak while ice descending from the nunatak flows in the opposite direction. The resulting convergence generates a zone of strongly reduced horizontal velocity and weak vertical advection. In this region, modeled basal velocities are on the order of millimeters per year, implying residence times of hundreds of thousands to millions of years within the stagnation zone. Basal ice recovered from the ALHIC1902 borehole suggests that the basal ice was stagnant and remained so possibly from the start of the most recent glaciation in the Allan Hills [Shackleton et al., 2025].

A simplified two-dimensional analytical solution captures the essential dynamics of this configuration. In this idealized representation, a break in slope in the glacier bed produces a stagnation point where horizontal velocities approach zero (Fig. B.1). Although the analytical model neglects the 3D geometry, nonlinear ice rheology, ice crystal properties and other details of the real system, it reproduces the key dynamical

feature observed in the numerical simulations: the development of extremely slow basal flow near the base of the nunatak where ice flowing from different directions converges.

3.4.2 Imported preservation by advection

Not all old ice in the Allan Hills BIA is preserved through local stagnation. The CDS site, represented by the ALHIC2301 core, contains ice older than 5 Ma despite relatively rapid modeled basal velocities. Ice flow into the CDS is driven by a strong surface slope and thickness gradient, producing basal velocities roughly forty times larger than those observed near the stagnation zone. If the age of ice were controlled solely by local residence time, these higher velocities would be expected to produce younger ice.

Instead, geochemical observations indicate that the CDS ice originates from a different upstream environment. The ALHIC2301 core exhibits distinctly lower $\delta^{38}\text{Ar}/^{36}\text{Ar}$ and $\delta^{18}\text{O}$ values than the other Allan Hills cores for ice younger than 1.5 Ma. Because $\delta^{38}\text{Ar}/^{36}\text{Ar}$ is related to firn thickness in the accumulation zone, and $\delta^{18}\text{O}$ reflects the temperature of snow formation, these isotopic differences imply that the CDS ice formed under different accumulation conditions. The most parsimonious interpretation is therefore that the CDS ice was formed elsewhere and subsequently advected into the BIA.

This mechanism explains why the CDS site deviates from the relationship between basal flow speed and maximum ice age. At sites near the nunatak, older ice corresponds to slower modeled basal flow, indicating a zone of persistent ice stagnation. In contrast, the CDS site preserves old ice that reflects its upstream origin (likely higher in elevation) rather than the residence time of ice within the stagnant ice flow regime.

3.4.3 Inferences about long-term ice flow variations

The Allan Hills ice cores commonly exhibit highly deformed and discontinuous stratigraphy, including folded and overturned layers [Shackleton et al., 2025; Kirkpatrick et al., 2025]. Several dynamical mechanisms have been proposed to explain folding in glacial ice. Recirculating flow structures such as Moffatt eddies can develop in viscous flow near sharp topographic depressions [Meyer & Creyts, 2017] and recumbent folds can develop through the progressive rotation of stratigraphic disturbances in a shearing ice column [Jacobson & Waddington, 2005].

Our results suggest an additional mechanism related to changes in ice thickness relative to local topogra-

phy. When the nunatak is exposed, downslope flow from the bedrock opposes regional ice-sheet motion and produces the stagnation zone identified upstream of the ALHIC1902 and ALHIC2302 boreholes. When the ice surface is higher, however, flow may instead pass over the topography, as illustrated by the submerged ridge transect shown in Fig. B.4. These two regimes are both observed in the modern ice sheet but at different locations, providing a space-for-time analogy for how flow near the nunatak might respond to changes in ice thickness over glacial time scales. Alternation between these flow regimes through glacial–interglacial cycles could locally reorganize the strain field experienced by the ice column and contribute to the folded and discontinuous stratigraphy observed in Allan Hills ice cores.

3.4.4 Implications for the exploration for multimillion year ice

Our results suggest that identifying future sites containing very old ice requires consideration of both dynamical regimes identified here. Regions of extremely slow basal flow created by topographic convergence remain promising targets because they can preserve ice locally for millions of years. By extrapolating the relationship shown in Figure 3.3a to all observed to slower modeled velocities, we show that the Allan Hills BIA may contain >10 Ma ice (Fig. 3.3b). At the same time, very old ice may still exist where present-day velocities are relatively large indicating a confluence of mechanisms operating to preserve old ice.

High-resolution ice-flow modeling combined with geochemical tracers therefore provides a powerful approach for identifying potential oldest-ice sites and interpreting the climate records preserved in the ice cores. By linking ice-core geochemistry and modeled flow dynamics, the framework developed here can be applied to other Antarctic BIAs to distinguish between locally preserved and advected old ice. Because these same flow regimes also control the surface stranding of meteorites [Benoit, 1995], such modeling may additionally help refine search areas for meteorite recovery by identifying zones where ice flow and ablation favor long-term surface concentration.

Taken together, our observations demonstrate that the Allan Hills BIA preserves ancient ice through two distinct dynamical pathways. Most boreholes containing very old ice are associated with zones of slow or stagnant basal motion produced by topographically induced flow convergence upstream of the nunatak. In contrast, the Cul-de-sac preserves old ice that was transported from a distinct upstream source region and later exposed through ablation. This dual-mode preservation mechanism reconciles the spatial distribution

of ice-core ages, ice-flow velocities, and geochemical signatures across the Allan Hills region and provides a more general framework for interpreting the dynamics of multimillion year ice preservation.

3.5 Methods

3.5.1 Ice flow model

We develop a finite element 3D full-Stokes thermo-mechanically coupled ice flow model using the Firedrake [Ham et al., 2023] Python package to characterize the ice flow in the AH BIA. We used continuous piecewise quadratic basis functions for velocity and piecewise linear for pressure, i.e. the Taylor-Hood element. We used a Newton line search solver from PETSc [Balay et al., 2025] for the nonlinear system.

3.5.2 Governing equations

We construct the momentum balance in the variational form,

$$\int_{\Omega} [\tau \cdot \epsilon(v) - q \nabla \cdot u - p \nabla \cdot v - \rho g \cdot v] dx$$

where

$$\epsilon(v) = \frac{1}{2} (\nabla v + \nabla v^T)$$

Here, ρ is the density of ice, u is the velocity, p is pressure, and v and q are test functions. The initial solve uses a Newtonian stress-strain relationship defined as $\tau = 2\mu_0\epsilon(u)$ where μ_0 is the initial viscosity given as a constant of $\mu_0 = 5 \cdot 10^{15}$ Pa·s. To solve for the initial temperature field, we construct the variational form of the diffusion-advection equation with a geothermal flux term supplied at the bed as

$$\int_{\Omega} k \nabla T \cdot \nabla \phi - \rho c T u \cdot \nabla \phi dx - \int_{\Gamma} \dot{s} \phi ds_b$$

where c is the specific heat capacity, k is the thermal diffusivity, $\dot{s} = 50 \text{ mW m}^{-2}$ is the prescribed geothermal heat flux, and ϕ is a scalar test function.

Using the initial solve for the velocity and temperature fields, we solve for a non-linear, temperature-

dependent viscosity field. First, we define the effective strain rate as

$$\epsilon_{\text{eff}} = \sqrt{\frac{1}{2} \left\{ \epsilon(u) \cdot \epsilon(u) + \text{tr} [\epsilon(u)]^2 \right\}}$$

where $\epsilon(u)$ is the strain rate tensor. We then calculate the temperature-dependent rate factor A given in Cuffey & Paterson [2010]

$$A(T) = A_0 \exp \left[-\frac{Q}{R} \left(\frac{1}{T + 273.15} - \frac{1}{263} \right) \right]$$

We solve for a new viscosity field following a Glen's Flow treatment such that

$$\mu = \frac{1}{2} \cdot E \cdot A(T)^{-\frac{1}{n}} \epsilon_{\text{eff}}^{\frac{1}{n}-1}$$

where E is the enhancement factor (see Fig. B.1). We then iteratively solve for a new flow field, temperature, and viscosity until the percent change in the residual value of the previous viscosity field to the new viscosity field stabilizes.

3.5.3 Boundary conditions

A coarse resolution mesh domain is created that encompasses most of the local accumulation zone and the drilled AH BIA ice core boreholes. The coarse mesh horizontal resolution is 800 m in the horizontal and we extrude 5 layers in the vertical dimension. We perform a full Stokes 3D solve on the coarse mesh with no flow on the vertical faces and the bed. On the surface, we allow for free slip. The solution to the coarse solve is then interpolated onto a smaller, fine mesh with a horizontal resolution of 80 m that includes the ALHIC1901, ALHIC1902, ALHIC2201, ALHIC2301, ALHIC2302, and ALHIC2401 boreholes. We chose not to include ALHIC1903 in the domain of the fine mesh as ALHIC1903 was not drilled to bedrock and does not contain old ice. The interpolated flow field is used as the flow boundary condition for the vertical faces of the fine mesh domain. We allow free slip at the ice surface. The basal velocity boundary condition is set to zero ice flow.

For the surface temperature condition, we define a surface temperature of -33°C at 2000 m elevation

and vary the temperature across the surface of the domain according to the dry adiabatic lapse rate or 1°C per 100 m of elevation change. We initially model a steady state temperature profile given a geothermal heat flux of 50 mW m^{-2} . We run a time-dependent surface temperature boundary condition hand-tuned to 2 K of warming over the last 150 years. The modeled ice profile temperature is scaled by a site-dependent constant offset that provides the best fit at each borehole to account for changes in local mean temperature not accounted for by the dry adiabatic lapse rate. The average temperature residual between the observed borehole temperature and the modeled temperature with the local offset applied is 0.04 C. This rate of warming, 2 K of warming over 150 years, is of similar rate of warming found in previous reconstructions of East Antarctic warming over a similar time period [Steig et al., 2009]. Additionally, a geothermal heat flux of 50 mW m^{-2} is consistent with the seismically-derived average geothermal heat flux of the East Antarctic of 52 mW m^{-2} [Shen et al., 2020].

3.5.4 Statement on the use of AI

The authors declare the use of generative AI in the research and writing process. According to the GAIDeT taxonomy (2025), the following tasks were delegated to GAI tools under full human supervision:

- Code generation
- Proofreading and editing
- Adapting and adjusting emotional tone
- Reformatting

We also used GAI to derive mathematical equations for stagnation flow in a wedge.

The GAI tool used was: ChatGPT (GPT-5.x). Responsibility for the final manuscript lies entirely with the authors. GAI tools are not listed as authors and do not bear responsibility for the final outcomes.

Chapter 4

Passive seismic reflections observed at the Eastwind Glacier grounding line, Antarctica

In preparation for submission to Geophysical Research Letters

4.1 Chapter Abstract

The Antarctic Ice Sheet discharges ice through glacier grounding lines—the transition between grounded ice and the floating ice shelves. Grounding line position and stability influence the ice sheet response to future climate scenarios. Yet, investigations of the glacier grounding zones using in situ instrumentation remains limited. In this study, we perform a passive seismic survey of the Eastwind Glacier grounding zone using distributed acoustic sensing. We recorded 12 hours of seismic data and create a catalog of 61 icequakes related to crevassing and basal stick-slip motion near the grounding line. Icequakes are modulated by tidal displacement with a cluster of activity occurring at a crevasse field near the end of the fiber optic array. We observe, for the first time, seismic reflections off of the grounding line from an icequake source. This study showcases the power of DAS for investigating glacier grounding lines in Antarctica.

4.2 Introduction

The grounding line is the transition zone from grounded ice to floating ice shelves. Ice shelves and grounded ice are dynamically linked at the grounding line [Wiens et al., 2024] thus the grounding line plays an important role in modulating discharge of grounded ice from the Antarctic Ice Sheet (AIS). Ice shelves flex due to the rise and fall of the tides causing the grounding line to migrate [Vaughan, 1995; Tsai & Gudmundsson, 2015] that causes the grounding line to act more like a "grounding zone". Changes in either the ice sheet or ice shelf components can influence the grounding zone stability with feedbacks to both the ice shelf and ice sheet geometry [Schoof, 2007]. While many of the East Antarctic Ice Sheet (EAIS) grounding lines have remained stable over the past 30 years in large part due to their prograde beds, their tenuously stable geometries do not preclude them to indefinite stability in future warming scenarios [Rignot et al., 2026; Sergienko & Haseloff, 2023]. However, the geometric shape and the material composition of the sediments or bedrock at Antarctic grounding zones remain largely unknown. Additionally, understanding how these properties change in space and time necessitates new geophysical observations.

Here, we leverage this persistent icequake activity at an EAIS grounding zone to probe the subglacial characteristics and dominant dynamical processes through high-resolution sampling of the seismo-acoustic wavefield. Changes in the stress balance across the grounding zone can promote fracture [Hulbe et al., 2016; Millan et al., 2022; Le Bris et al., 2025; Podolskiy & Walter, 2016] or stick-slip motion [Lucas et al., 2023] generating icequakes. Ice fractures have been observed at the grounding zone of many of the polar ice sheet outlet glaciers [Hulbe et al., 2016; Millan et al., 2022] indicating icequake activity at grounding zones is ubiquitous. Previous work has linked icequake activity to spatially varying bed conditions [Smith, 2006]. Efforts to model grounding line dynamics indicate basal friction and the chosen basal friction law plays a significant role in the projected ice mass discharge across grounding lines [Brondex et al., 2017]. Processes at the grounding zone not only act to destabilize the grounding line position, but can also stabilize it. For example, sedimentation beneath ice shelves at the grounding zone has been shown to stabilize grounding line retreat [Alley et al., 2007]. In order to identify the relative contributions of each of these dynamical processes to grounding zone destabilization, the basal environment at the grounding zone must first be characterized. To do this, we exploit frequent icequake events and subsequent seismic wave propagation at the grounding zone to "image" the grounding line location and bed conditions at the Eastwind Glacier,

Antarctica grounding zone.

Eastwind Glacier is ~ 40 km from McMurdo Station and is an advantageous study site for its accessibility and logistical importance. Eastwind Glacier flows off of Mount Terror and contributes to the McMurdo Ice Shelf (MIS) [Hoffman et al., 2025] (see Fig. 4.1). The MIS is susceptible to warm water intrusion into the sub-ice-shelf cavity that melts the underside of the ice shelf and causes grounding line retreat [Robinson et al., 2010]. Changes in the stress balance and tides can cause calving at the MIS terminus [Banwell et al., 2017]. While MIS is currently stable, surface melting, basal melting, or retreat of the terminus could alter the stability of the grounding line in the future, thus continued geophysical surveying is critical [Hill et al., 2022]. Fiber optic cables have proven to be robust sensing elements in glaciated environments where other geophysical instrumentation may be difficult or costly to deploy [Manos et al., 2024]. Distributed acoustic sensing (DAS) exploits the Rayleigh backscatter of light in optical fiber to measure strain, strain rate, or phase continuously and at high spatio-temporal resolution [Lindsey & Martin, 2021]. DAS is sensitive to glacier microseismicity and capable of locating events with similar accuracy to conventional geophone arrays [Hudson et al., 2021]. We present results from a two day passive seismic DAS survey of the Eastwind Glacier grounding zone. We identify clusters of icequake activity attributed to lateral shear margins and crevasse fields at the grounding zone. Icequakes originating from the floating ice shelf reflect off of the grounding line providing a new way to interrogate the position of glacier grounding lines and the basal sediment characteristics at the glacier grounding zone.

4.3 Materials and Methods

4.3.1 Distributed Acoustic Sensing (DAS)

We utilize a previously deployed singlemode fiber optic cable during the 2023-24 field season [Agnew et al., 2025b] that was well coupled at the time of this study given an accumulation rate of 0.7 ma^{-1} [Hoffman et al., 2025]. Figure 1 shows the cable geometry and location of the DAS interrogator at the start of the fiber optic cable (A in Fig. 4.1). We used a Sintela ONYX™ DAS interrogator with a pulse rate of 2 kHz with a spatial resolution and gauge length of 6.38 m. The fiber optic cable is approximately 2.2 km long and crosses the grounding line at multiple points. Since the cable is only sensitive to longitudinal strains [Benioff, 1935],

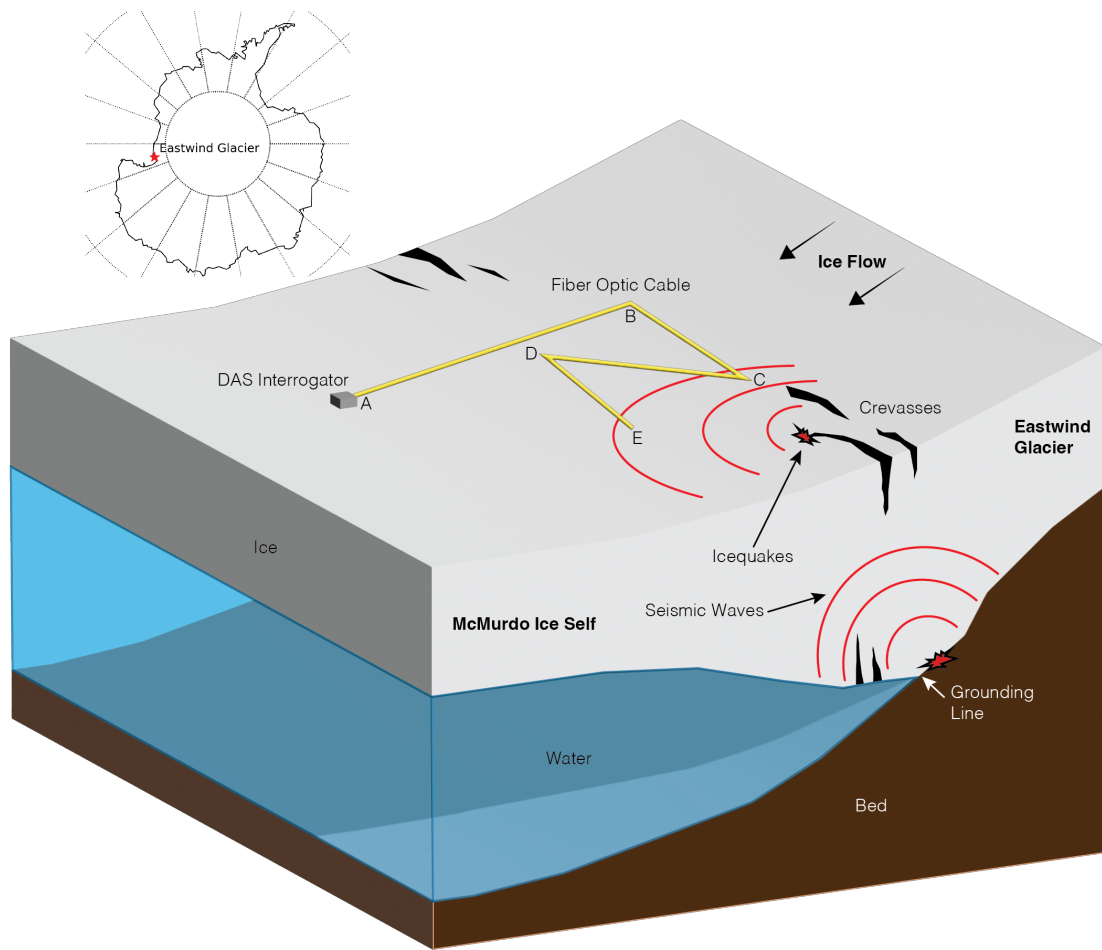


Figure 4.1: A schematic of the Eastwind Glacier grounding line and its location in Antarctica. The orientation of the fiber cable and location of the DAS interrogator is shown relative to the glacier grounding line. Labels **A-E** denote the ends of or turns in the fiber cable array. The fiber cable is buried ~ 1 m in snow and intersects the grounding line at multiple points along its path.

it is advantageous to incorporate turns in the cable to make determining seismic event locations possible. A Honda 2 kW generator was used to power the DAS system and the DAS was kept in a tent for protection from the environment. Generator noise is visible on the first ~ 10 channels; however, it attenuates rapidly and does not affect the rest of the array.

We recorded two different periods of passive seismic data over the course of two days during the 2024-25 Antarctic field season. Our first observation period began on December 18, 2024 at 04:54 UTC and ended at 11:19 UTC of the same day. Our next observation period began on December 18, 2024 at 23:07 UTC and ended on December 19, 2024 at 05:16 UTC totaling 12 hours of data. We recorded on 351 channels with a spatial resolution of 6.38 m. The distance of fiber over which the phase change of backscattered light is averaged for each channel is called the gauge length which was set to 6.38 m. Pulse rate, analogous to sampling frequency, was set to 2 kHz for the duration of the experiment and files are saved in 1-minute increments.

4.3.2 Tide Data

We use the CATS2008 model [Padman et al., 2008] for ocean tide height beneath MIS and near the Eastwind Glacier grounding zone (see Fig. C.1. While this is an inverse model and not derived from any locally deployed sensors, it is a good approximation for the the timing and magnitude of tides where getting accurate tidal height measurements is impractical.

4.3.3 Detecting icequakes

The DAS data is first bandpass filtered between 2 Hz and 500 Hz which encompasses the frequency range of most cryoseismic sources of interest [Podolskiy & Walter, 2016] and normalize each channel waveform. Next, we apply an FK filter the data between 1500 ms^{-1} and 7000 ms^{-1} to remove instrument noise and preserve signals traveling within the range of the apparent body and surface wave velocities in ice.

We use ObsPy and its recursive STA/LTA (short-term average/long-term average) algorithm [Beyreuther et al., 2010] to detect events on the DAS timeseries data. Icequakes are typically low in magnitude (e.g. ~ -1.0 [Walter et al., 2010] thus many signals of interest are close to the noise floor of the DAS data making STA/LTA less effective. To improve the signal-to-noise (SNR) ratio, we first perform phase-weighted stack-

ing on 5 channel segments of the DAS data which suppresses incoherent noise before running the STA/LTA on the stacked channels. Detections across the cable that are coincident in time are grouped together and classified as an icequake. We identify 81 events using our automated detection workflow and remove 20 events from the catalogue after manual inspection as these events were cases of instrument noise resulting in a catalog of 61 positively identified icequakes. Following detection, we make manual event arrival picks on each DAS channel by picking the arrival of the first completely resolvable wave phase across the array to mitigate ambiguity in picks. When possible, we pick the first arriving wave phase such as the P or S wave, but occasionally surface waves are the only resolvable wave phase and are picked.

4.3.4 Locating icequakes

To find the most probable event location, we perform a Markov chain Monte Carlo (MCMC) ensemble sampling [Foreman-Mackey et al., 2013] of the parameter space and model the arrival times at each channel given as

$$t_{arrival} = t_{event} + \frac{distance}{wavespeed} \quad (4.1)$$

where $t_{arrival}$ is the time of the first arrival picked manually, t_{event} is the time of the seismic event, and distance is the distance from the channel to the seismic event calculated from sampling possible x , y , and z coordinates of the event location. Residuals between the picked arrivals and the modeled arrivals using a log likelihood function as defined by Foreman-Maxkey et al. 2013. The ensemble sampler is initialized with 64 walkers and each walker with 5000 steps before removing the first 500 steps to account for sampler burn in. Waveforms are sorted by distance to the event such that we solve for the true wave speed rather than apparent wave speed as observed by the DAS cable.

4.3.5 Modeling waveforms

We create a two-dimensional (2D) acoustic wave equation solver using finite elements in the *Firedrake* Python package. The synthetic wavelet is generated from the observed signal of the dominant wave packet, most likely the Rayleigh wave, traveling at $\sim 1910 \text{ ms}^{-1}$ to model a synthetic icequake source. The domain consists of a two homogeneous media: one that approximates the ice shelf with a velocity of ~ 1910

ms^{-1} and one that approximates bedrock beneath the grounded ice, shown in Agnew et al. [2025b], with a wave velocity of 3000 ms^{-1} , separated by a linear grounding line hand tuned to fit observed reflections. The grounding line geometry used for the model is similar to the grounding line found by remote sensing [Gerrish et al., 2025] (see Section Text C.1 for more model construction details).

4.4 Results

4.4.1 Spatiotemporal variability and clustering of icequakes at the grounding line

Icequake clusters are not uniformly distributed across the grounding zone, but are centered on three general locations—the most active being near a crevasse field at the grid-eastern shear margin (Fig. 4.2b). During the beginning of the observation period and coincident with a rising tide (see Fig. C.1), icequake activity is highest on the ice shelf side of the grounding line. Significant basal crevassing has been observed at the Eastwind Glacier grounding line with crevasses growing size towards the ice shelf [Hoffman et al., 2025]. Enhanced icequake activity during rising tides at the ice shelf may be do to ice shelf tidal flexure and associated tensile stresses coupled with the injection of seawater into the newly opened crevasses during a rising tide [Christianson et al., 2016; Begeman et al., 2020]. However, activity is highest during the falling tide (Fig. C.1), which is consistent with the findings of others [Barruol et al., 2013; Le Bris et al., 2025] and suggests that tidally induced icequakes are ubiquitous across different grounding line geometries and characteristics.

4.4.2 Seismic waves from an icequake reflect off of the grounding line

DAS offers the ability to recover high quality recordings of icequakes at the grounding line not possible with traditional seismic sensors. We highlight a single icequake (star in Fig. 4.2) plotted both in optical distance and distance from the best fit location from the MCMC (see Section 4.3). Approximate wave speeds of $\sim 3610 \text{ ms}^{-1}$, $\sim 2000 \text{ ms}^{-1}$, and $\sim 1860 \text{ ms}^{-1}$ are plotted on top of the waveforms in Fig. 4.3b and are consistent with the range of possible wave speeds of P, S, and Rayleigh waves traveling through ice, respectively [Booth et al., 2020]. Because of the spatially dense measurements and increased waveform coherence made possible by DAS, we can identify clear seismic reflections indicated by negative relative

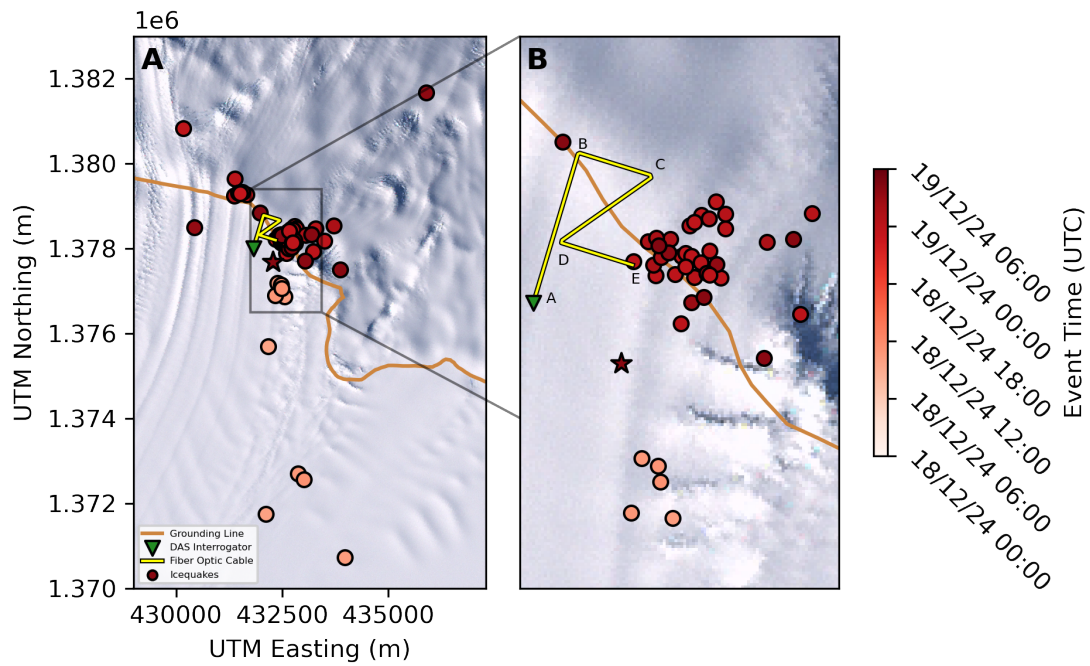


Figure 4.2: a) All quality controlled icequake detections and located icequakes during the observation period. b) Zoom in near the grounding zone that highlights the region of most frequent icequake activity. Background image is from the Landsat Imagery Mosaic Antarctica (LIMA). The star indicates the icequake of interest shown in 4.3 and further discussed in Section 4.4.2

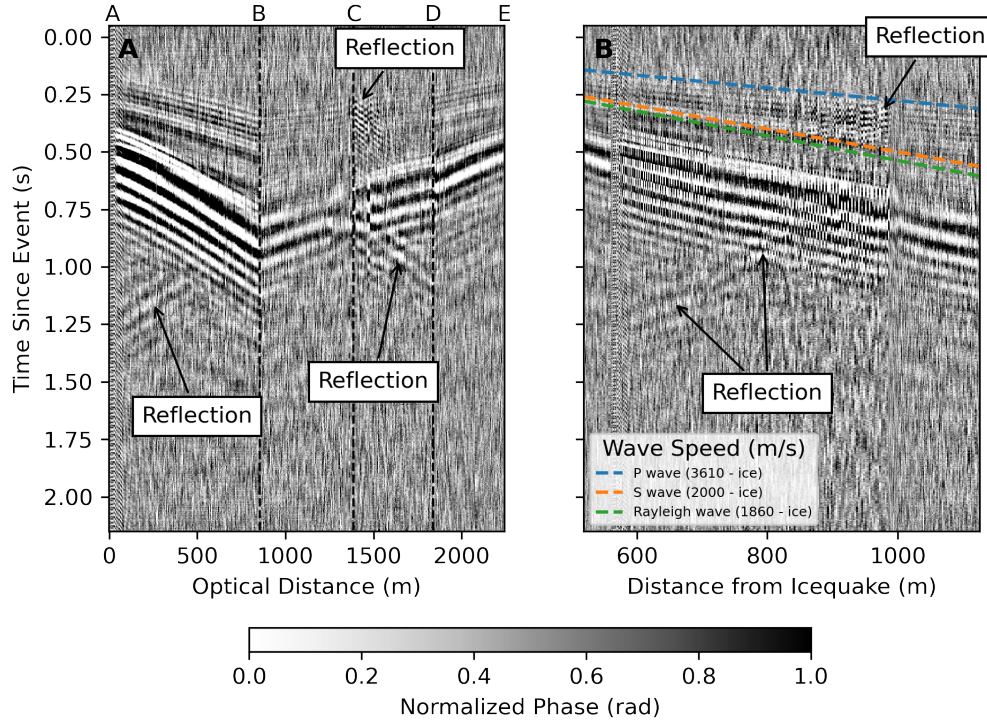


Figure 4.3: Highlighted icequake (star in Fig. 4.2 plotted in **a**) optical distance with 0 being nearest to the DAS interrogator and **b**) distance from the icequake that shows distinct reflections indicated by the opposite traveling seismic waves. Label A-E in **a**) indicate the ends of the cable (A and E) and any turns (B-D) that are also shown in Figs 4.1 and 4.2. Approximate wave speeds for the P wave ($\sim 3610 \text{ ms}^{-1}$), S wave ($\sim 2000 \text{ ms}^{-1}$), and Rayleigh wave ($\sim 1860 \text{ ms}^{-1}$) plotted in **b**) are consistent with waves traveling through ice.

wavenumbers (waves traveling in the opposite lateral direction, Fig. 4.3).

The 2D geometry of the fiber optic array aids in identifying likely source interfaces for the reflection. The reflected waves are visible on cable sections A-B and C-D, the sections of cable with components perpendicular to the grounding line. The sections parallel to the grounding, B-C and D-E, are insensitive to the reflected wave (Fig. 4.3a) indicating that the reflections are originating from the direction of the grounding line. To our knowledge, this is the first observation of an icequake source producing reflections from an Antarctic ice shelf grounding line, the most likely interface for seismic reflection given the direction of travel of the reflected wave (see Fig. 4.3a).

4.5 Discussion

Reflections from icequakes typically occur at the ice-bed interface where a strong velocity contrast is often present [Walter et al., 2009, 2020; Hudson et al., 2020] and imaging the ice-bed interface using reflection seismology is most often done using active seismic sources [Agnew et al., 2025b,a]. However, passive seismology for "imaging" the grounding line paves the way for revealing temporal links in grounding zone change and monitoring grounding line migration. To test the potential for passive seismology as a tool for monitoring grounding zone dynamics, we first validate model grounding line reflections and compare to our observations. Then, we calculate the grounding line reflection coefficient, which can be used to differentiate a range of possible materials at the ice interface causing the reflection.

4.5.1 Simulated acoustic wave propagation at the grounding zone

One possible source of reflections from icequakes are open crevasses. Eastwind Glacier has significant basal crevasse that grows in size towards the edge of the ice shelf, but the basal crevasses do not start until ~ 2 km downstream of the grounding line [Hoffman et al., 2025]—well outside the aperture of the fiber array. Moreover, limited surface crevassing on the main flow path onto the ice shelf was present at time of the seismic survey [Agnew et al., 2025b; Hoffman et al., 2025]. If significant crevassing was present near the grounding line, we should expect to detect more icequakes located at the grounding line in the main ice flow than are detected (see Fig. 4.2). Instead, icequakes primarily cluster in the crevasse fields on the shear margins or out on the ice shelf where basal rifting has been observed [Hoffman et al., 2025] suggesting little crevasse activity at the grounding line.

Next, we create a 2D finite element model to test the horizontal acoustic wave propagation of an example icequake source (star in Fig. 4.2) and compare to observations. We use the dominant wave velocity found in Fig. 4.3 of $\sim 1910 \text{ ms}^{-1}$ for the ice velocity and a velocity of $\sim 3000 \text{ ms}^{-1}$ for bed material at the grounding zone. Modeled scalar acoustic wave propagation is shown in Figure C.2 with coherent reflections at the grounding line crossing the fiber optic array (Fig. C.2c). Furthermore, the waveforms of the observed icequake and reflections compared to the modeled waveforms match closely (Fig. C.3) with deviations likely caused by the simplistic grounding line geometry and the longitudinal sensitivity of the DAS fiber cable array. The specularity of the reflections in both the observed waveforms and modeled waveforms

point towards a large relatively linear interface most consistent with the grounding line.

4.5.2 Ice-grounding zone interface reflection coefficient

The material properties of the sediment or bedrock at grounding zones across Antarctic are largely unknown. However, the geometry and the material make-up (i.e. bedrock, dilatant till, or unconsolidated sediments among others) of the grounding zone controls the flow of ice onto the ice shelf [Alley et al., 2007] thus it is critical to have geophysical observations at the grounding zone to accurately model ice sheet discharge. Typical methods for recovering the reflection coefficient of the ice-bed interface are through active seismic surveys and amplitude variation with angle [Zechmann et al., 2018]. However, active seismic surveys are expensive and logistically complicated to carry-out. Here, we investigate if retrieving a reasonable reflection coefficient at the grounding line is possible using icequakes as the source.

We limit our analysis to the linear portion of the cable oriented roughly orthogonal to the grounding line (labeled A-B in Fig. 4.1) to get the best sensitivity to reflected waves. First, incident and reflected waves from the candidate icequakes are separated using a directional sensitive FK-filter. The wave amplitudes are then calculated as the maximum absolute wavelet amplitude of the incoming (incident) and outgoing (reflected) waves at each channel excluding the end channels to prevent artifacts from FK-filtering and instrument noise (Fig. 4.4). The reflection coefficient R is calculated as $R = \frac{A_r}{A_i}$ where A_r is the amplitude of the reflected wave and A_i is the amplitude of the incident wave. Icequakes that produce median reflection coefficients $R > 1$ are removed. The mean reflection coefficient found is $R = 0.19$, which if we assume average incident angles $< 45^\circ$, is most consistent with an ice-consolidated till interface [Zechmann et al., 2018]. If the reflections were primarily from crevasses, either air-filled or water-filled, we would expect a much larger average reflection (i.e. $\sim 0.2 - 1$).

4.6 Conclusion

We observe seismic wave reflections off of an Antarctic Ice Sheet grounding line using a dense distributed acoustic sensing fiber array deployed across the grounding zone. Icequakes are primarily located in the grid-eastern shear margin with icequakes causing reflections primarily located on the floating ice shelf side of the grounding line. The observed wave propagation direction and decrease in reflected wave amplitude

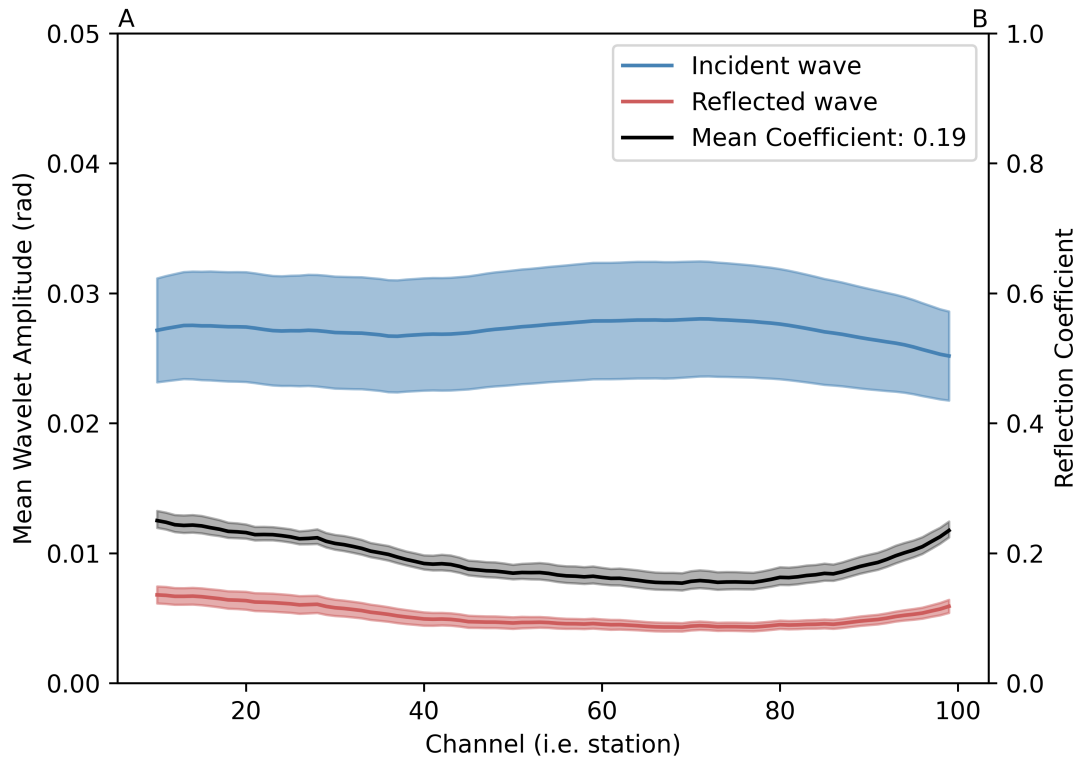


Figure 4.4: Reflection coefficients computed from maximum absolute wavelet amplitude at channels in the straight section of cable (labeled A-B in Figures 4.1 to 4.3) orthogonal to the grounding line. Assuming the average angle of incidence is $<45^\circ$, the interface is most consistent with ice and consolidated till [Zechmann et al., 2018]

most closely match an ice-consolidated till interface at the grounding zone. Utilizing icequakes for imaging the grounding zone unlocks the possibility for monitoring changes of the grounding zone through time, a critical constraint on future ice sheet discharge projections.

Chapter 5

Conclusion

Collectively, the goal of this work is to reveal dynamical processes important for ice flow and, in particular, contributing to the spatiotemporal variability in melt and ice discharge using dense fiber optic sensing arrays. I explore the capabilities of distributed optical fiber sensing (DOFS) in both mountain glacier and polar ice sheet environments. I show how two DOFS techniques: distributed acoustic sensing (DAS) and distributed temperature sensing (DTS) can be used to recover high-resolution spatiotemporal temperature and acoustic wavefield measurements. Not only do I test the feasibility of implementing DOFS technology in field-based glaciological research settings, I also add new observations into ice flow and wavefield models. Through the use of novel methodologies in fiber optic sensing, I reveal new insights in the spatiotemporal variability of glacier melt and ice sheet flow.

In Chapter 2, I measure the ambient supraglacial seismo-acoustic wavefield using a DAS fiber optic array deployed along the flowline of Rhonegletscher, Switzerland. The soundscape of the glacier surface in the ablation zone is dominated by the sounds of melt and turbulent runoff. The accumulation zone, however, is quieter owing to the lack of runoff and the snowy surface as opposed to the icy ablation zone surface. It has been shown that amplitude of the downstream of a glacier reflects the proglacial river discharge [Podolskiy et al., 2023], but whether this relation holds true for the glacier surface is unknown. I show, here, that the seismo-acoustic wavefield densely measured by DAS can be used to monitor meltwater runoff and thus proglacial discharge. Machine learning models trained on the DAS recorded seismo-acoustic wavefield can not only infer discharge better than models trained on meteo data alone, but can quantify the variability in

discharge. Furthermore, I show that the section of cable with the most significant sensitivity to discharge is the section located at the snow line, which indicates that DAS can be used as a spatially-distributed supraglacial melt sensor.

In Chapter 3, as part of the National Science Foundation Center for Oldest Ice Exploration (NSF COLDEX), I model the flow of ice in the Allan Hills Blue Ice Area (BIA) where the oldest ice ever recovered on Earth has been found. However, the ice cores are discontinuous reflecting a complicated flow history. Without constraints on ice flow in the Allan Hills, interpretation of the paleoclimate records preserved in the ice cores is difficult. I create a finite element thermo-mechanically coupled three-dimensional ice flow model validating the results using surface velocity measurements from GNSS ablation stakes and borehole temperature measurements from DTS. I show that multiple flow regimes exist within the Allan Hills BIA: stagnation flow caused by remnant ice on a nearby nunatak strands old ice at the ALHIC1902/2301 core site, ice in the Cul-de-sac is advected from upstream, and the ice flowing through the ALHIC1901/2201 core site is forced up and over a subglacial ridge analogous to ice flow overriding the Allan Hills nunataks. I show that older ice is found in areas of progressively slower flowing basal ice as indicated by the ice flow model. By integrating measurements of trace gases in the ice cores, I find that the ice in the Cul-de-sac has a distinct origin than the ice from the main ice flow into the ALHIC1901/2201 and ALHIC1902/2301 core sites. This work provides a generalizable framework for combining geochemical tracers and ice flow models to search for old ice in other locations and BIAs across the Antarctic continent.

In Chapter 4, I investigate the dynamical processes of ice flow across the grounding zone. The grounding zone controls the flux of grounded ice into the ocean ultimately contributing to sea level rise. In conjunction with field training with team members from the NSF COLDEX and the NSF funded EGGS on TOAST project, I record icequakes at the Eastwind Glacier grounding zone using a previously deployed fiber optic cable instrumented with a DAS interrogator. During the course of this study, 61 icequakes are recorded and located to nearby crevasse fields in the shear margin and in the main ice flow on the ice shelf. Icequake event frequency is modulated by ice shelf tide displacement altering the stress balance in the grounding zone and within crevasses inducing icequakes. Because of the high spatial resolution and high signal-to-noise ratio, it is possible to resolve coherent reflections from the grounding line. Moreover, I find an average reflection coefficient of 0.19, indicative of an ice-consolidated till [Zechmann et al., 2018]. This work paves

the way for using DAS to monitor spatiotemporal variability of the mechanical properties at the bed of the grounding zone. With constraints on the grounding zone mechanical properties and stresses, more accurate models describing ice flow at the grounding zone and projections of sea level rise can be made.

5.1 Perspectives

Models of ice melt, discharge, or flow are fundamentally limited by the quality of geophysical observations on which the models rely. It is impossible to assess the accuracy or the validity of a model without measurements of the variables that are of interest in the system that is being modeled. However, making observations of the dynamically processes of glaciers and ice sheets is difficult. I argue here that by using DOFS technology, observations of the mechanics of ice motion, fracture, and melt can be made more simply and without trade-offs between spatio-temporal resolution and aperture. In this dissertation I demonstrate the utility of DAS and DTS individually, but greater insights can be gained by deploying collocated DAS and DTS fiber cables [Gräff et al., 2025]. One of the areas most limited in geophysical observations and process understanding is that of the polar ice shelves and associated grounding zones.

Glacier and ice sheet adjustment is typically considered to be a low-pass filter of the climate responding only to the long term climate [Roe, 2011]. Yet, ice shelves can disintegrate rapidly over the course of a single melt season [Banwell et al., 2013] causing the upstream ice to speed up in response [Scambos et al., 2004]. This rapidly collapse originally implicated supraglacial meltwater ponding as the culprit, but the recent disintegration of the Conger-Glenzer Ice Shelf suggest alternative dynamical processes [Walker et al., 2024]. One of the biggest challenges facing glaciology community today is reconciling these competing models of glacier and ice sheet response to climate change with the observational record [Bassis et al., 2024]. Producing models that can accurately predict future ice sheet or ice shelf stability thus sea level rise rely on integrating the short term elastic behavior of the ice shelf-grounding zone with the long term trend of the polar ice sheets. Large uncertainties in our projections in the stability of the polar ice sheets and and sea level rise currently exist due, in part, to our poor understanding of these processes. Dense geophysical observations at the on the ice shelf and at the grounding line, possibly using DOFS but not exclusively, however, can reveal the coupling of stress imbalances from hydrologic or atmospheric forcings to the dynamics occurring at the grounding zone. Ultimately, these unique observations will be used to

inform models of sea level rise and, furthermore, reduce our uncertainty in the response of the cryosphere to climate change.

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Chapter A

Appendix 1: DAS to Discharge: Using Distributed Acoustic Sensing (DAS) to infer glacier runoff

A.1 Hyperparameter tuning.

Figure A.1 shows the hyperparameter tuning of model input window sizes and a comparison between model types. The window sizes correspond to the number of 30 second time steps to include in the window for model training, validation, and testing. For example, a window size of 20 corresponds to a window size of 10 minutes time by number of channels. The linear machine learning model used for comparison performs an order of magnitude worse in all tested window sizes. The Long Short-Term Memory (LSTM) model and Neural Network (NN) models perform similarly; however, LSTM tends to be the better performing model with less variance over most of the tested window sizes hence our selection for further testing. Models with large window sizes (>500) tend to break down in their predictive precision and accuracy as the number of

training batches is reduced. Models with small window sizes (<20) tend to perform significantly better and have lower variance which may be explained by the increase in training batches and that the discharge value is relatively stable at sufficiently small timescales. To combat these artifacts that misrepresent the power of machine learning inference of discharge, we opted to choose a window size sufficiently large enough to capture the meltwater transport down stream to the discharge gauge which we chose as a window size of 200 corresponding to 100 minutes of data. A learning rate of 0.001 was chosen for the analysis as we found varying the learning rate slightly did not change results significantly and a learning rate of 0.001 is fairly standard in machine learning applications.

A.2 Data shuffling and splitting for model ingest.

To include all possible surface melt regimes encountered during the observation period in our analysis, we performed 100 iterations of data windowing, shuffling, model creation, and batching into training, validation, and testing data sets. The ratio of training, validation, and test data set size was kept fixed between each iteration at 70, 20, and 10 percent, respectively. The windows are non-overlapping and the data sets split into training, validation, and test sets are completely separate from each other for each iteration. Depending on the batch size, which was kept at 32 for the study but allowed to vary during the window size testing phase, the test set would contain inconsistent batch sizes. In these cases, the final batch created from the remainder time steps would be removed from the test data set as to keep consistent batch sizes for model ingest and prediction.

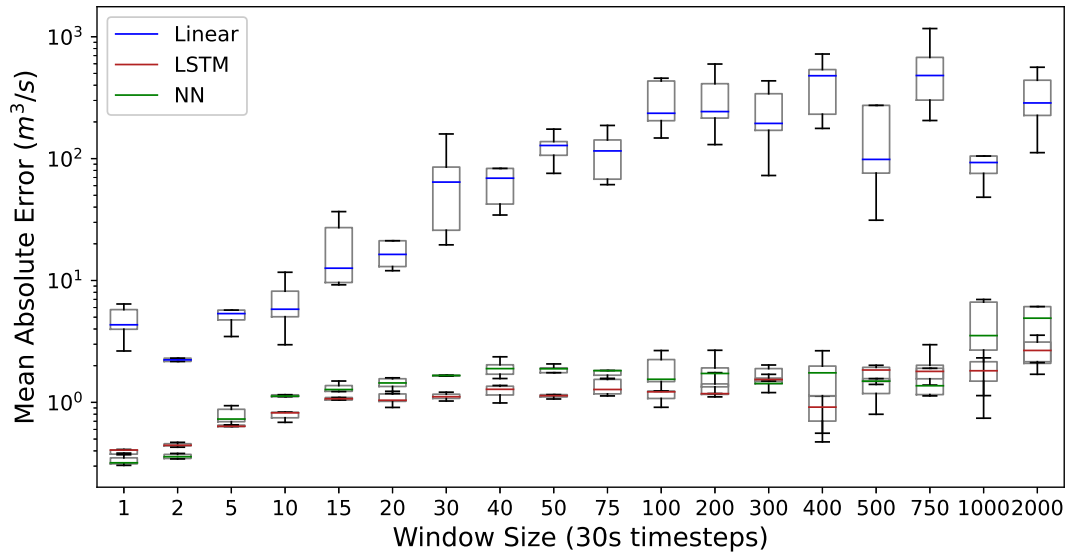


Figure A.1: Mean absolute error (MAE) from 18 window size hyperparameterizations tested with three different types of machine learning model architectures: Linear, LSTM, and NN. The box signifies the interquartile range and the whiskers are the last data point in the range that are less than 1.5 times the interquartile range. A linear machine learning framework performs significantly worse than the LSTM and NN architectures. LSTM and NN architectures perform comparably well, however, the LSTM achieves better performance at a wider range of window sizes.

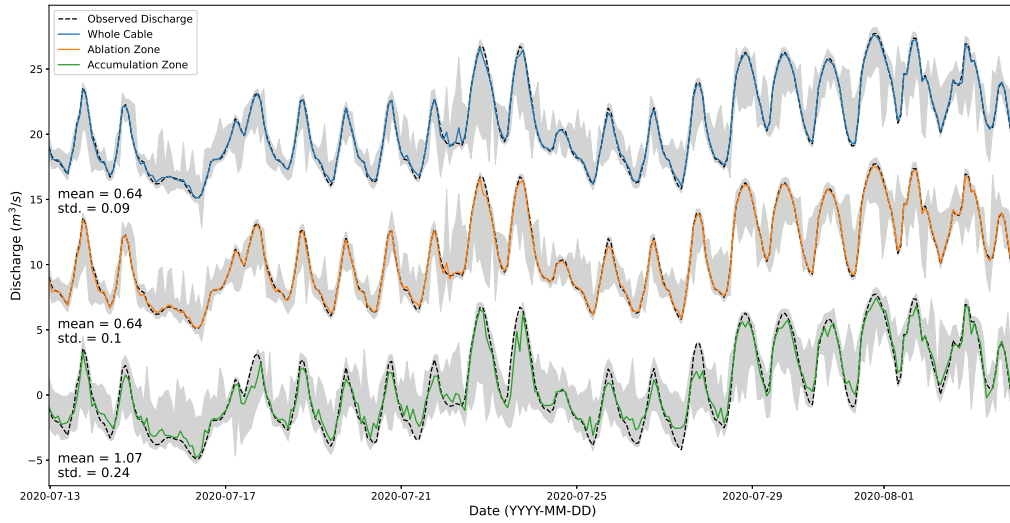


Figure A.2: Model performance from 100 model runs for models trained on all data (top, blue), data taken only from the section of cable in the ablation zone as determined to be from channels 0 - 1602 (middle, orange), and data taken only from the section of cable in the accumulation zone as determined to be channels numbered higher than 1602 (bottom, green). The mean and standard deviation (std) is calculated for the mean absolute error for each of the 100 model run sets and reported below the respective predictions on the sectioned data sets. Note that the blue and green curves are systematically offset from each other along the y-axis by $\pm 10 \text{ m}^3/\text{s}$ for comparison purposes.

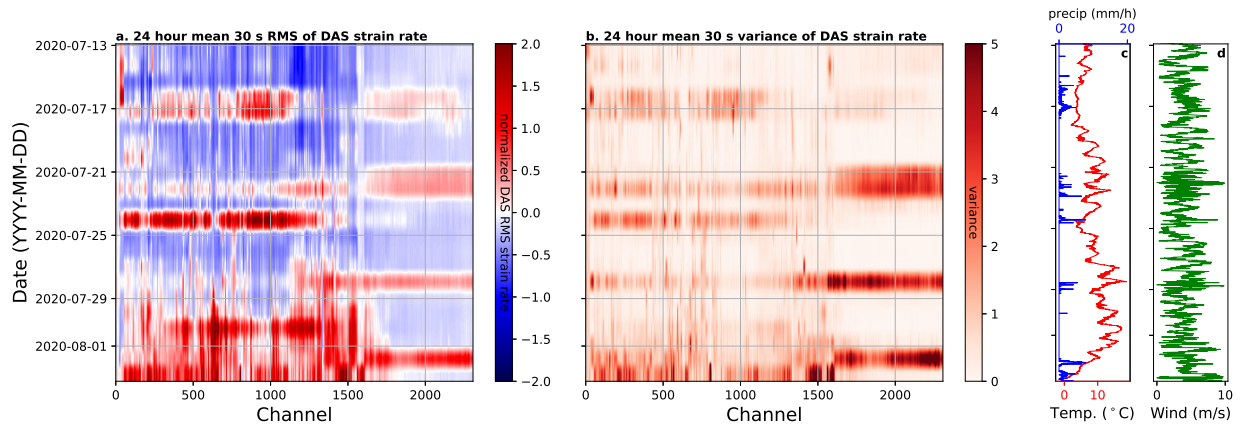


Figure A.3: **a)** Mean of normalized 30 s RMS strain rate over 24 hour periods with respect to channel. **b)** Variance of normalized 30 s RMS strain rate over 24 hour periods. **c)** Hourly temperature, precipitation, and **d)** wind data from 10 min recordings at Grimsel Hospitz meteo station (Swiss Federal Office of Meteorology and Climatology MeteoSwiss).

Chapter B

Appendix 2: Present-day ice flow at sites preserving old ice in the Allan Hills Blue Ice Area (BIA), Antarctica

B.1 Supplementary text

B.1.1 GNSS velocity measurements

A network of 54 ablation stakes was deployed during the 2023-24 field season spanning from the ablation zone to the local accumulation zone in Allan Hills, Antarctica. Stakes were placed in a diamond shape about one ice thickness apart around phase-sensitive radar (ApRES) acquisition sites along multiple ice flowlines. Additional stakes were placed in the vicinity of the ALHIC2301 and ALHIC2401 drill sites to constrain ice flow speed and direction into this area of interest. In the ablation zone, ablation stakes were cut to ~ 1.5 m long three-quarter-inch steel conduit drilled into blue ice where possible ~ 50 cm deep. In the accumulation

zone, stakes were cut to ~ 2 m and drilled into the snow surface ~ 1 m deep.

In both the 2023-24 and 2024-25 field seasons, we surveyed stake positions in ~ 15 min durations using a NetR9 Trimble GNSS receiver (1 s sampling rate) mounted on a PVC tube that could fit over the metal conduit stakes. A GNSS base station on the slow-flowing (< 0.2 m/yr) blue ice recorded data during both survey periods. Files were converted to RINEX 2.11 format using the Trimble Convert to RINEX utility. We obtained daily base station positions and uncertainties from the Canadian Spatial Reference System Precise Point Positioning (CSRS-PPP) static processing tool. Individual stake positions were calculated in Emlid Studio with static processing of the measurement and base station RINEX files, using the CSRS-PPP base station locations to correct for satellite signal errors. We take the stake position uncertainties to be the sum of uncertainties in the Emlid Studio and CSRS-PPP processing.

Surface velocities were calculated by differencing the stake positions and dividing by the interval between measurements (approximately 1 year). Because of the need to resolve small displacements, we only accept stakes with two years of FIX solutions (solutions with centimeter-level precision) for these calculations. In six cases where the Emlid Studio workflow yielded a lower-precision solution, we obtained a stake location through direct CSRS-PPP static processing of the measurement data. Surface velocities observed from GNSS observations captured within the bounds of the high-resolution model centered on the ablation zone are reported in Table B.1.

B.1.2 Ice borehole temperature measurements

Ice temperature profiles were measured in seven AH boreholes. The ALHIC1901 borehole temperature in 3 consecutive seasons starting in the 2022-23 field season. ALHIC1902, ALHIC1903, ALHIC2201, and ALHIC2301 were measured during the 2023-24 season. ALHIC2301, ALHIC2302, and ALHIC2401 were measured during the 2024-25 season. All boreholes are air-filled and are either 4 inches or 10 inches in

diameter.

To measure ice temperature, we use distributed temperature sensing (DTS) which measures the Stokes and anti-Stokes Raman backscatter in fiber optic cables. The ratio between the anti-Stokes and Stokes backscatter amplitude is proportional to temperature of the glass fiber cable. The fiber optic cable is constructed with two multimode and two singlemode loose-tube-buffered fibers surrounded by 4 tight-buffered singlemode fibers. The two loose-tube-buffered multimode fibers are spliced together on one end to form a "U-bend" configuration such that two spatially-coincident sections of optical backscatter are measured along the optical path—one down-going and one up-going. We report the borehole temperature at each timestep as the average between these two coincident sections of optical fiber. Double-ended collections were performed to precisely constrain differential attenuation and mitigate loss in signal fidelity across bad splices or physical connectors. We post-process the DTS data using the *dtscalibration* Python package des Tombe et al. [2020].

Temperature measurements were taken a minimum of ~ 24 hours after cable deployment to ensure thermal equilibrium of the cable with the borehole temperature. Temperature is measured every 0.255 meters along the optical pathway. Individual temperature measurement duration and total observation length varied between cable deployments; however, we take the average temperature measurement of measurement durations to obtain a single borehole temperature profile for each borehole measured for each season. In the 2022-23 field season, an external Pt100 temperature probe was used to calibrate a spooled section of fiber contained within an insulated cooler at the surface. We did not use an external probe to calibrate the temperature measurements in the 2023-24 or 2024-25, rather we used the internal temperature sensor and reference fiber section to do the temperature calibration. The temperature offset in the absolute temperature profile in the ALHIC1901 borehole between the 2022-23 and 2023-24 seasons (see Fig. B.2e) is likely due to this change in methodology; however, the residuals of the borehole temperature profiles from a linear

trend (inset Fig. B.2k) show little change between seasons indicating good relative temperature accuracy.

B.1.3 Analytical solution for stagnation flow

We begin with the classical 2D stagnation flow solution in the upper half-plane $y > 0$, where the streamfunction is given by $\Psi(x, y) = Axy$. This corresponds to a velocity field:

$$u = Ax, v = -Ay$$

Which, in polar coordinates may be written as

$$\Psi(r, \theta) = Ar^2 \sin(2\theta)$$

valid in the domain $0 < \theta < \pi$. This solution essentially corresponds to a flat bed topography. When the bed contains a slope connected to a flat region, the solution can be found by applying the conformal mapping,

$$\zeta = z^{\alpha/\pi}$$

which maps the upper half-plane (a wedge of angle π) to a wedge of angle α . Under this transformation, the streamfunction transforms to,

$$\Psi(r, \theta) = Ar^{2\pi/\alpha} \sin(2\pi\theta/\alpha)$$

This streamfunction satisfies the biharmonic equation and yields velocity components consistent with incompressible, steady Stokes flow. The stagnation point remains at the origin, and the glacier bed is a stream surface.

The velocity components of the analytical solution are plotted in Supplemental Figure B.4.

B.1.4 Ice core geochemistry

Argon dating, $\delta^{18}\text{O}$, and $\delta^{38}\text{Ar}/^{36}\text{Ar}$ measurements presented here follow prior work at the Allan Hills. We include previously published Yan et al. [2019]; Shackleton et al. [2025]; Marks-Peterson et al. [2026]; Davidge [2025] and new $\delta^{18}\text{O}$ measurements from ALHIC2201, ALHIC2301, and ALHIC2302. $\delta^{38}\text{Ar}/^{36}\text{Ar}$ is measured as part of the ^{40}Ar sample age. $\delta^{18}\text{O}$ is either measured on the residual meltwater from ^{40}Ar age samples or, where these data were not available, on the drilling chips from the core section where an ^{40}Ar age sample was collected. We present average values from ice <1.5 Ma. Ar measurements were made at Princeton University following methods described in Bender et al. Bender et al. [2008], Yan et al. Yan et al. [2019] and Shackleton et al. Shackleton et al. [2025].

Water isotope data were analyzed at the University of Maine or the University of Washington using either a Picarro L2130-i or a L2140-i cavity-ring-down laser spectrometer following previous studies Schauer et al. [2016]. Analyses proceeded with four reference waters included after every 10th sample. The last three of six total injections were averaged for each sample and reference-water vial. The reference waters cover a spread of Antarctic values (created from excess Holocene and glacial ice from West Antarctic and South Pole)Steig et al. [2013, 2021]; these reference waters were previously calibrated to the VSMOW-SLAP water isotope scale. Sample calibrations were based on a three-point linear regression, with the fourth reference water treated as an unknown and used for quality control. Data are reported as deviations (δ) in ‰ from VSMOW. Precision and accuracy (one standard deviation), based on the quality control reference water, was <0.6 ‰ for δD and <0.1 ‰ for $\delta^{18}\text{O}$.

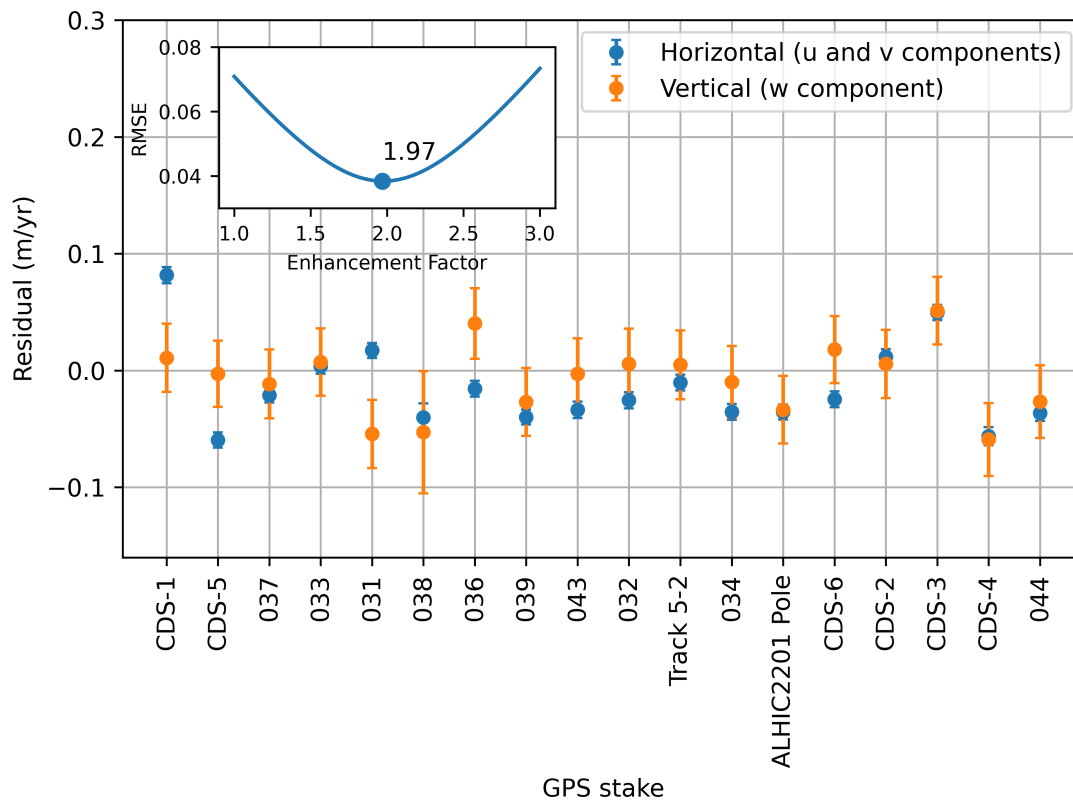


Figure B.1: Model residuals from GNSS surface velocity measurements in the horizontal speed (u and v components) and the vertical velocity. Error bars are cumulative errors between subsequent year GNSS observations. The figure inset shows the optimal enhancement factor to produce the lowest model residuals from the GNSS surface velocity measurements.

name	x (m)	y (m)	u (m yr ⁻¹)	v (m yr ⁻¹)	w (m yr ⁻¹)	horizontal uncertainty (m)	vertical uncertainty (m)
CDS-1	508405.2180	-1352365.5240	-0.1427	-0.0187	-0.0366	0.0069	0.0292
CDS-5	507813.3948	-1352291.6200	-0.0509	0.0462	0.0089	0.0065	0.0284
37	510523.6074	-1354288.9900	-0.0719	-0.0498	0.0225	0.0064	0.0294
33	510640.4039	-1353604.6280	-0.0406	-0.0524	0.0084	0.0064	0.0288
31	511441.4203	-1353039.0680	-0.0707	-0.0746	0.0368	0.0065	0.0292
38	510286.1781	-1354899.4520	-0.0702	-0.0349	0.0625	0.012	0.0523
36	511324.0963	-1353726.8450	-0.0853	-0.1232	-0.0194	0.0068	0.0302
39	510429.5550	-1354897.5090	-0.0715	-0.0536	0.039	0.0064	0.0291
43	510358.1820	-1354970.9160	-0.0650	-0.0414	0.0122	0.007	0.0303
32	512006.7780	-1353842.5950	-0.1079	-0.1797	0.0009	0.0069	0.0302
Track 5-2	511728.5274	-1353432.1450	-0.0957	-0.1391	-0.0139	0.0067	0.0294
34	511206.7230	-1354409.1260	-0.0511	-0.1557	0.0231	0.0068	0.0308
ALHIC2201 Pole	510357.2208	-1354869.7070	-0.0646	-0.0425	0.0459	0.0064	0.0289
CDS-6	507865.8740	-1353083.1190	-0.0071	0.0328	-0.0129	0.0067	0.0288
CDS-2	508167.7890	-1352727.3690	-0.1392	-0.0557	0.0374	0.0065	0.0292
CDS-3	508058.1091	-1352262.5690	-0.1261	-0.0012	0.0391	0.0065	0.029
CDS-4	508063.2626	-1352622.9310	-0.2146	0.0006	0.1084	0.0077	0.0312
44	510356.8367	-1354821.9610	-0.0650	-0.0487	0.0393	0.0068	0.0311

Table B.1: GNSS position velocity measurements of ablation stakes placed during the 2023-24 season. Positions were recorded during the 2023-24 and 2024-25 field seasons.

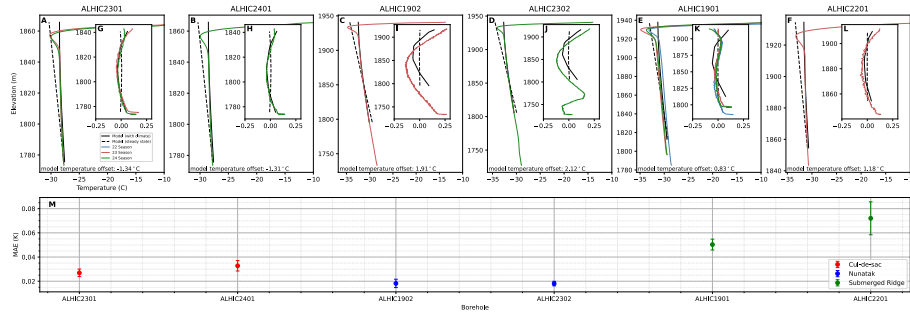


Figure B.2: Absolute ice profile temperature (a-f) measured in accessible Allan Hills boreholes (colored line) compared to modeled ice profile temperatures (black solid). For context, we show a temperature model that has no time-dependent climate trend applied to the boundary and is indicative of a steady state simulation (dashed line). The residual from a linear profile (g-l) shows the contribution of conduction and advection to the observed and modeled temperature signal. A constant temperature offset calculated as the mean temperature residual below the seasonal signal is applied to each modeled borehole temperature measurement to improve model fit. A 2 K linear warming trend over a 150 year period was used to create ice temperature profiles with a climate signal applied. We fit a temperature model with no advection component (only considering time-dependent conduction) to observed borehole temperatures (m) and calculate the mean absolute error (MAE) to show the role of advection. Boreholes that have better fits or a lower MAE are less influenced by ice flow in describing their observed temperature profiles.

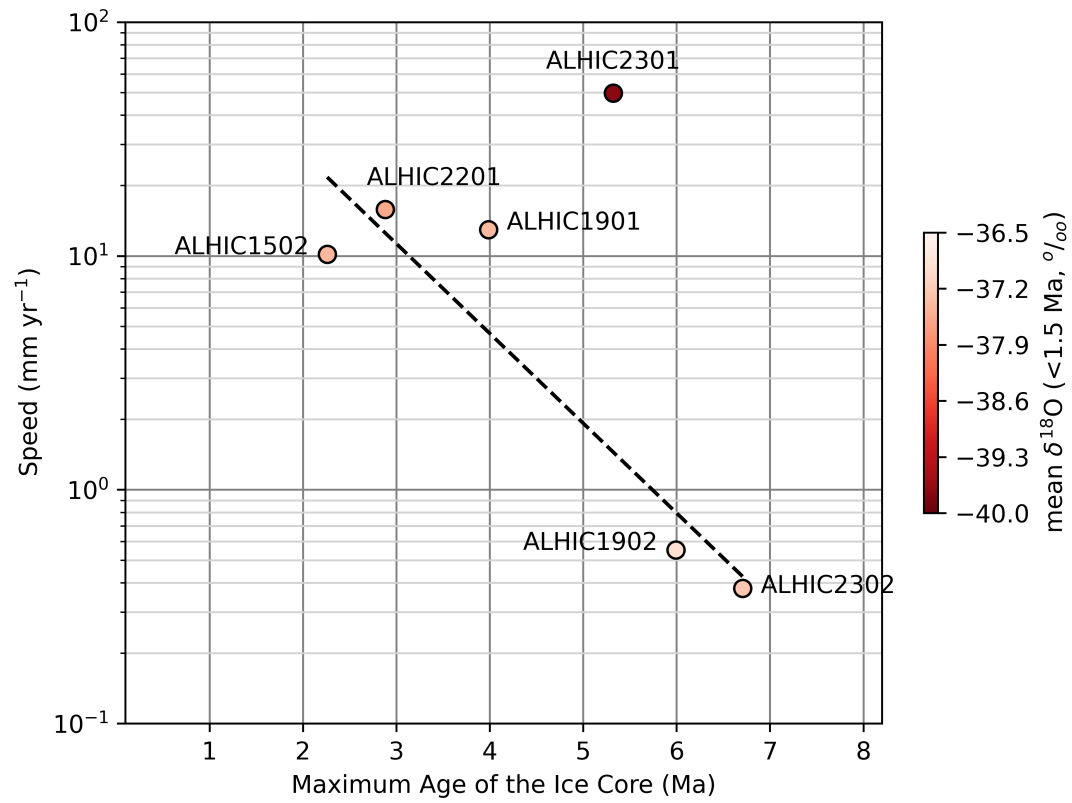


Figure B.3: Similar to Figure 3.3a, but now colored by the mean $\delta^{18}\text{O}$ ratio.

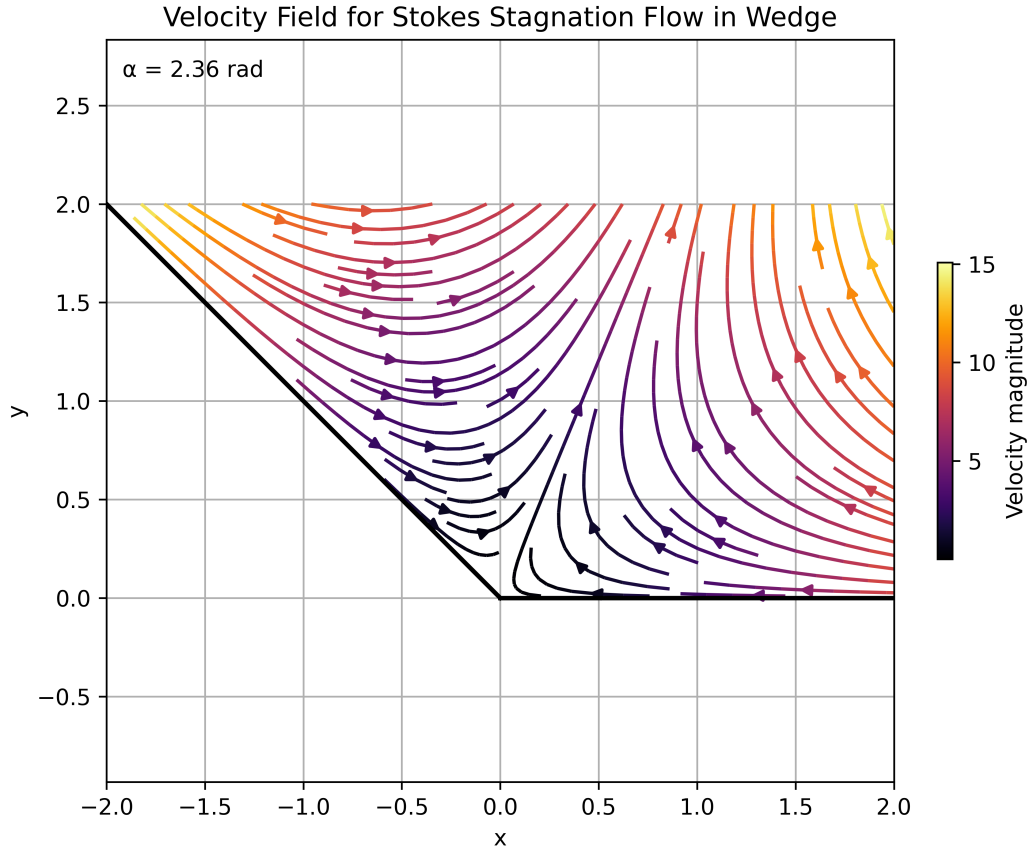


Figure B.4: Cross-sectional profiles of ice velocity in the idealized analytical solution for Stokes flow. The x (horizontal) and y (vertical) axes are nondimensional, as is the velocity magnitude scale. The 2D solution shown here predicts strong vertical ice flow, whereas in 3D ice is able to escape in the third (into and out of the page) direction.

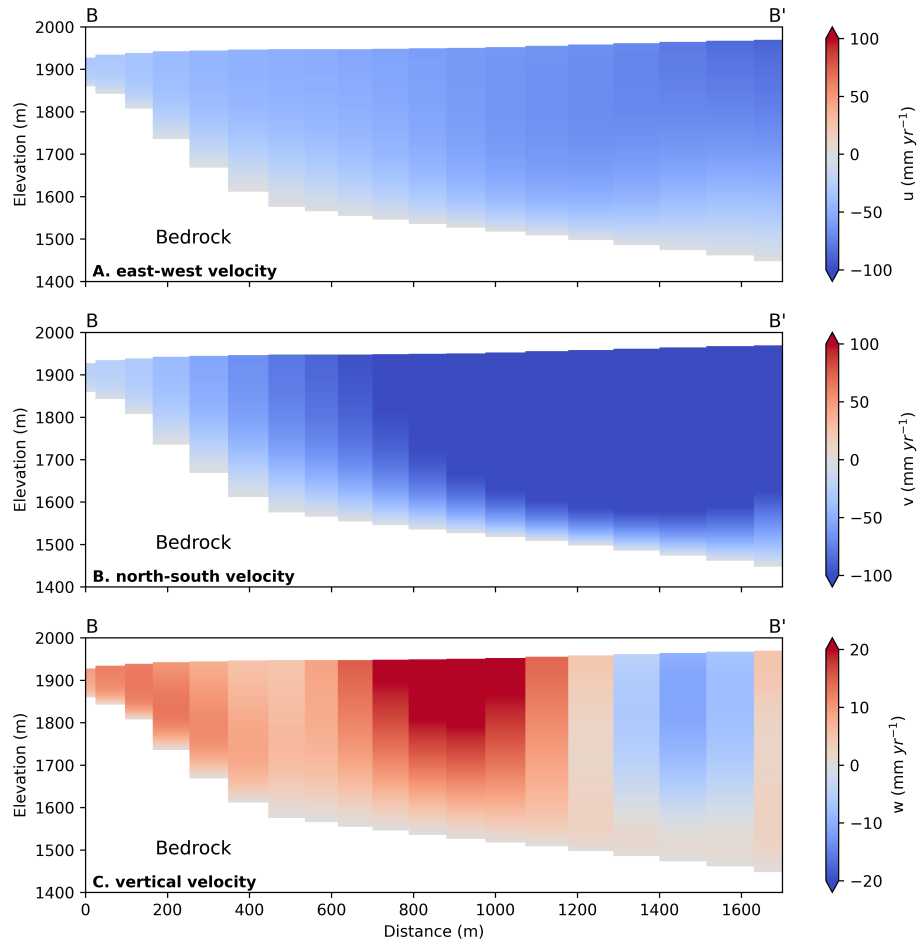


Figure B.5: As in Figure 3.2 in the main text, except taken for a flow line through the AL-HIC1901/ALHIC2201 boreholes.

Chapter C

Appendix 3: Passive seismic reflections observed at the Eastwind Glacier grounding line, Antarctica

C.1 Two-dimensional wave propagation forward model

The two-dimensional scalar wave propagation forward model is solved using the finite element method coded in *Python* using the *Firedrake* software package Ham et al. [2023]. The scalar wave equation can be written as

$$\frac{1}{c^2} \frac{\partial^2 u}{\partial t^2} = \nabla^2 u$$

where u is some scalar field (e.g., pressure) and c is the wave velocity. The boundaries of the domain are absorbing (i.e. wave do not reflect off of the domain boundary). We use 3000 ms^{-1} for the seismic wave velocity of the basal material at the grounding zone and 1910 ms^{-1} for the seismic wave velocity of the ice.

For the source, we use a wavelet constructed from part of the observed waveform containing the dominant wave energy traveling at 1910 ms^{-1} (see Fig. 4.3). The model domain and wave propagation are shown in Fig. C.2.

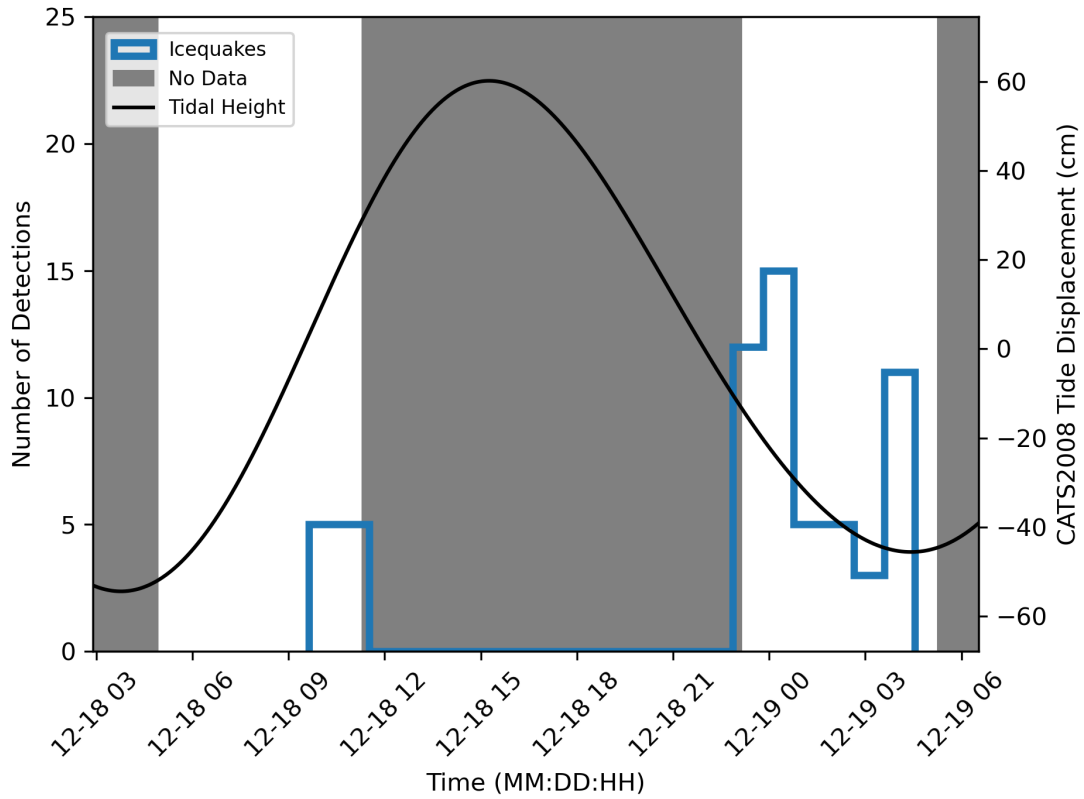


Figure C.1: Number of detected icequake events compared to the tide displacement at the McMurdo Ice Shelf. More icequakes occur during the falling tide than the rising tide.

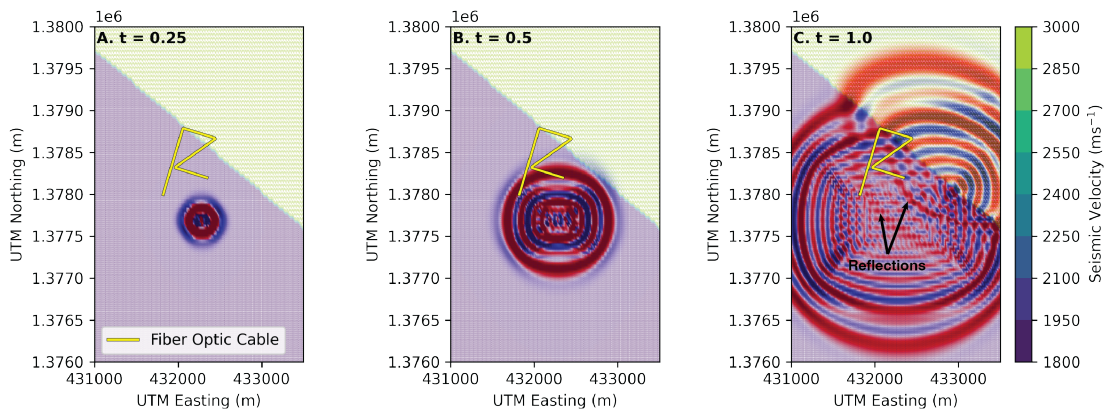


Figure C.2: Map view of the modeled waveforms of an icequake source at three time periods at a synthetic grounding line with similar geometry to the remote sensing derived grounding line at Eastwind Glacier, Antarctica [Gerrish et al., 2025].

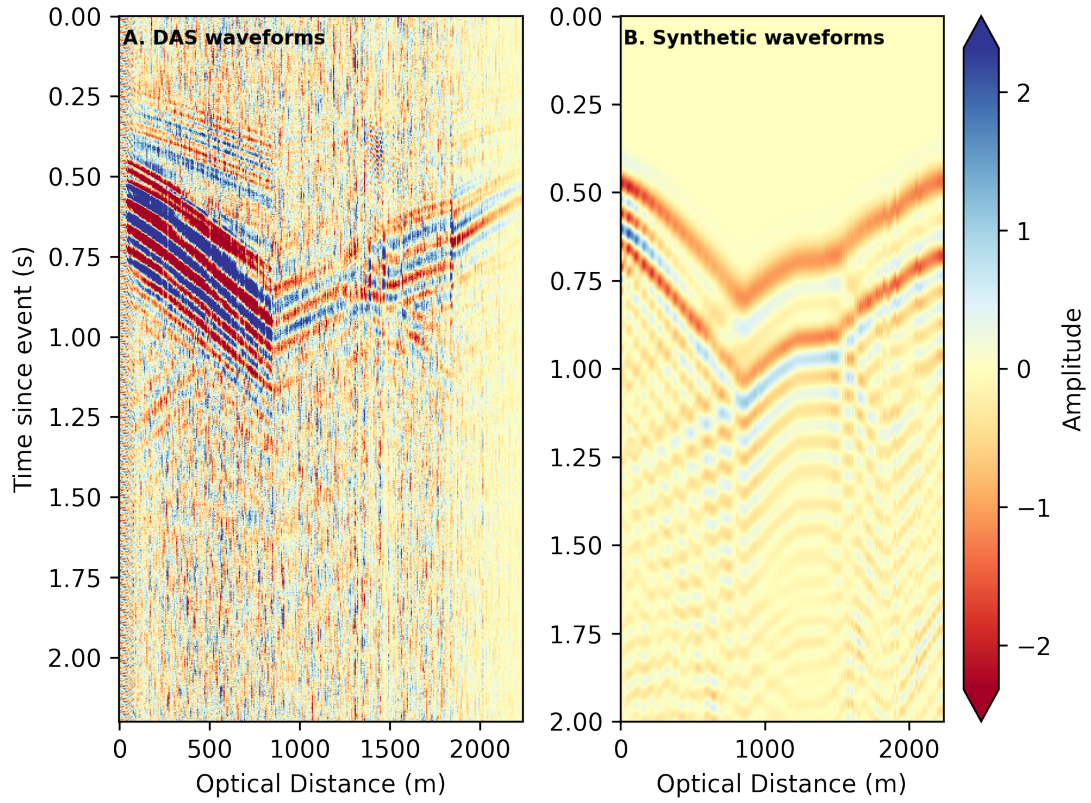


Figure C.3: Comparison of the **a)** observed icequake and reflection waveforms (same as Fig. 4.3a) and the **b)** modeled synthetic scalar waveforms. Observed and modeled waveforms qualitatively match closely with deviations likely due to imprecise grounding line geometry and the longitudinal sensitivity of the DAS fiber array.